ADVANCED LECTURES 1982 - THE CLIMATE OF THE WORLD

LECTURE 1 - Introduction D. E. JONES

The picture of a climatologist is usually of a man who produces great compendia of tables of average temperatures or pressures over the world and is responsible for all the obligatory climatological maps in the standard school atlases; otherwise he is the man who comes up with facts such as that last night was the coldest 20 January since 1926. Perhaps if we acknowledge that climatology is the study of climate or climates then it is easy to include as climatologists Charney who developed an explanation of the Sahel drought by a biogeophysical mechanism; John Mitchell and Manabe who are studying gigantic numerical models to determine how the physical parameters affect the climate in their models; the scientists taking ice cores in Greenland, examining tree rings in Canada or poring over old diaries and records extracting inexact information about the weather in the 13th or 17th centuries; and V. Bjertnes who studied large-scale atmospheric response to warming of the equatorial Pacific.

Climatology had obtained a bad name because the climatologists were content to be descriptive. Kendrew (1938) was a well known climatologist and actually writes in his book 'Climate':

> "In the study of climatology the primary interest lies in the facts and phenomena of the climates of the Earth in themselves, and in the natural environment of life. The investigation of the physical causes underlying the facts is an important as well as an interesting side of the study, but is to be regarded as subordinate"!

Nevertheless the compilers of descriptive studies are not to be decried there is no doubt that an essential requisite of any scientific investigation is to establish the facts - to obtain, validate and process the observational data so that the human mind can see what the problems are. Without the systematised data the numerical modellers of climate and of the general circulation (if there is a difference) could neither establish their models nor validate their results, and there is now a realisation that observational studies are essential to our understanding of how the atmosphere functions.

2. World Climate Programme

In 1974 the Executive Committee of WMO decided it was important for WMO to take a lead in promoting studies of climate changes and their impact on the natural environment of mankind and on food production. So, jointly with ICSU, the International Council of Scientific Unions, they arranged a high level scientific and technical World Climate conference in 1979 representing climate sensitive branches of national economies including agriculture, energy, water resources, fisheries and health. They reviewed knowledge of climatic change and variability and their implications for human activities. There followed an outline plan and basis for the World Climate Programme 1980-1983 (WMO 540).

WMO's role is to support and help co-ordinate activities, not to run them and they have set up a World Climate Programme Office within the Office of the WMO/ ICSU Joint Planning Staff (under Döös) at Geneva and have established a Joint Scientific Committee (JSC) with ICSU to provide, amongst other things, scientific guidance for the international aspects of the World Climate Research Programme to determine the main research objectives and to facilitate the exchange of information between scientists. The main objectives of the World Climate Programme may be summarised as follows: to improve knowledge of the natural variability of climate; to understand the underlying causes and mechanisms; to assess the sensitivity of climate to natural and man-made perturbations; and ultimately to predict future climatic changes; and to assist in the making of planning and investment decisions in climate-sensitive industries with a view to reducing the impact of adverse

climatic changes. The Programme will have three main elements: a climate data and applications programme; a programme to investigate the impacts of climate on human activities; a programme for research on climatic change and variability.

2. Climate Impact

You will notice that in the outline of objectives of the World Climate Programme the emphasis is on climate change and variability: this is natural as the great interest is really in the impact of such changes on man. Increasing world population and higher living standards have increased the pressure on natural resources of food, fibre, energy and water, and the balance between supply and demand could be seriously affected by even temporary changes in the climate of critical regions. The world's food supply is now perilously dependent on good weather. Abrupt changes in the length of the growing season, annual rainfall or mean temperature could be disastrous.

Climate impact is really outside the province of the meteorologist and very few studies have been made on the likely impact of marked or persistent changes of climate. There is no doubt that one of man's most natural energies is drought. A case in point is the Sahel, the area which separates the Sahara desert from Tropical West Africa and stretches from Senegal to the Sudan (Sahel is Arabic for border). It lies near the northern limit of the Intertropical Convergence Zone. There is normally rain in 4 months (summer months) amounting to 4-8 inches. It is natural pasture with short grass, perennial herbaceous plants, thorny shrubs and acacia and beobab trees. A long drought started in 1968 and by 1973 had led to the extinctio of virtually all crops and the loss of 60-70% of cattle. Loss of human life from starvation in 1973 alone was 100,000. The one-time nomadic people who migrated north and south with the ITCZ had been encouraged to become settled during the period when rainfall was relatively high. (An intriguing question is what would the authorities have done once the settling had taken place if the drought had been forecast even if it was forecast with certainty).

Drought occurs in many places from time to time even in such relatively wet places as the United Kingdom. Climate change is not necessarily all bad: a general rise of 2-3°C, as is quite possible as a result of increasing CO_2 in the atmosphere, could well lead to increased food production by prolonging the growing season in many areas, increasing the rainfall and enhancing photosynthesis; it would also produce considerable reduction in fuel consumption. However since the benefits and disadvantages will occur in different regions, and in spite of the great capacity of the modern world to transport enormous masses of food and other necessities there are undboutedly many millions of people who would be worse off. There is a great reluctance to transport people.

One of the functions of WCP is to increase world awareness of the available data and research for planners to use.

3. Climate Data

The range of observations required for all climate work is enormous, from the usual synoptic, climatological and upper air observations to radiation balance and stratospheric dust load. Coverage of the oceanic areas is particularly poor yet they account for two thirds of the earth's surface. Improved satellite observations giving world-wide coverage are an obvious advantage.

The major observational task has been defined by the World Climate Programme as the collection of a complete set of atmospheric, ocean and land-surface data for the period of at least the last 30 years. The FGGE (Global Weather Experiment) data set provides the most detailed picture to date of one annual cycle and is of great importance; another essential is the assembly of "long term instrumental record" data sets for the last 100 years or more.

The data-management planning to ensure this data is quality controlled and archived and made available internationally is prodigious, and the prime responsibility rests with NATIONAL archives.

4. What do we look to the Observations for?

The WCP define climate as "the synthesis of weather over the whole of a period long enough to establish its statistical ensemble properties (mean values; variances, probabilities of extreme events, etc) and is largely independent of any instantaneous state".

Obviously we use the observations to calculate the long period normals. Figure 1 is a typical climate field, but even here we must be careful to establish what periods the mean values are for, as many climatological fields show long term variability.

Other useful parameters are the inter-annual and shorter period variability and the extreme values. The observations will tell us whether the time variations have a coherence - obviously an important factor and this will be expanded on in a later lecture.

The Sahel sequence of droughts are a remarkable example of time coherence and quasi-periodicity - (see Figures 2-4 and Lockwood 1979, Faure and Gac, 1981, Boucher 1975, Nicholson 1981).

5. Empirical Studies

The atmosphere does most of its work on large geographical scales so that climate anomalies tend to be extensive in space - the Sahelian drought was longitudinally Africa-wide. On even larger scales it is common to find the variation of some element in one area correlated with its variation, or the variation of another element, in another area sometimes quite remote. This is of great interest because it indicates some processes are at work which may otherwise escape attention. (Hare 1979). Research has proceeded at a great pace in recent years as increasingly large data sets covering the world. Many statistically significant relationships have been found (teleconnections) but the period of reliable instrumental observation, especially in the upper air really is not long enough yet. There is only a limited number of statistical hypotheses that can be tested with a given amount of data.

Many examples of teleconnections have been demonstrated and some will be described in a later lecture by Mr Parker, eg connections of the southern oscillation, the Indian Monsoon, the el Niño ocean current and the upper trough over the eastern USA. Few of them have any significant predictive capabilities, and their man importance is the challenge they make to the theorist for they say something about the characteristic time and space scales of atmospheric behaviour, in many cases well ahead of our capacity to explain them.

Care has to be taken not to fall for the obvious answer. Research has demonstrated that the Sahel drought itself was caused by persistent anomalies of the general circulation, but this was not simply due to a southward shift in the subtropical high pressure belt (Flohn, 1977). Nor was the drought due to lack of water vapour. In fact at the height of the drought the high pressure zone was a little further north than usual, and the water vapour available for precipitation was above normal (Gal-Chen, 1977). First explanations interpreted these changes mainly as a function of the position of the ITCZ and subtropical high but this hypothesis remained controversial and more plausible explanations have been found after analyses of additional factors determining rain intensity.

It should be stressed that the ITCZ is not a belt of continuous uplift, but is usually made up of a number of cloud clusters separated by large expanses of clear sky. The cloud clusters are marked by heavy rainfall (>2 cm day^{-1}) and normally move westwards as easterly waves. Although an anomalously northern displacement of the ITCZ may be primarily responsible for abnormally wet years, the sub-Saharan drought appears to relate to factors governing the intensity of the rainy season (Nicholson 1981). The fact that the rainfall anomalies tend to be of the same sign to North and South of the Sahara, also strongly imply a general intensification of the Hadley cell during the drought and perhaps an expansion of its descending branch. Further evidence for this derives from increased rainfall in equatorial regions and an associated contraction of the ascending branch (Kanimitsu and Krishnamurti (1978)).

6. Long time series

Mr Parker will in his lecture mention the problems which arise in the monitoring of recent climate changes because of lack of long homogeneous (self-consistent) instrumental records, and Mr Gilchrist will discuss in some detail the use of proxy data such as the O^{18}/O^{16} ratio in ancient ice cores, and the width of tree rings, and what may be inferred about the past history of climate. Both types of work demand a tremendous effort with great problems of interpretation and of matching different records, and great care has to be taken not to infer too much from them.

Approaches to the Sahel problem include estimates of the level of Lake Chad back to about 1100 AD by synthesising historical, geological and polynological data, studies of famines in the past 300 years, and of course historical and instrumental records (since about 1900) of river flow (Faure and Gac, 1981).

7. Synoptic Climatology

There is some disagreement in the literature on the definition of the term "Synoptic climatology": most agree that it includes the description and analysis of the totality of weather (climate) of an area or region in terms of atmospheric circulation. The categories of circulation type have to be determined or defined, whether in terms of synoptic models such as the Bjerknes polar front cyclone or the tropical Easterly waves, or by means of more general concepts such as air flow or air mass type. Synoptic climatology is also concerned with the frequencies of the circulation types themselves and how the frequencies have changed over the years; the insights of synoptic climatology are also used for long range forecasting purposes. Synoptic Climatology is a major link between general circulation studies and climatology: to reproduce correctly the statistical distributions of the dynamical features such as depressions, strength of the westerlies, incidence of easterlies is a touchstone of a good general circulation model.

8. Climate Modelling

One of the ultimate goals of the World Climate Programme is climate prediction, and this can only be achieved on the basis of a sound quantitative theory of climate. This does not exist, but we may be able to develop such a theory if we begin to understand the relative importance of the various interacting physical processes that determine climate. Here climate modelling comes into its own, for we can modify and then "observe" the consequences.

There are two genres of model; one, the simulation models, started with Phillips (1956) and is now represented by highly complex models of for example Manabe and of the Dynamical Climatology branch which was referred to in the Advanced Lectures on the General Circulation. The others are simple one-dimensional (or occasionally two-dimensional) models expressing a few simply coupled mechanisms and taking as imposed conditions all the other components of climate (Sellars 1969). The simple models help to provide an understanding of the coupled feed-back processes that are too complicated to interpret in the simulation models. Dr Carson will say more about these simple models.

Dr Slingo has already described atmospheric circulation models; these are of course the same as climate models except that the latter inevitably need to express the coupling between the oceans and the atmosphere and in fact to include the circulation of the ocean itself.

Whereas the general circulation modeller is interested explicitly in the dynamics of the atmosphere, examining eddy transports and vertical cross-sections of atmospheric flow, the climate modeller is particularly interested in resulting surface phenomena such as sea surface temperature, and rainfall. The interest is indicated by the fact that Manabe and Holloway (1975) use Köppen's classification of surface climates to get a general impression of the skill of their climate simulations. The climate modeller needs to run models for many years rather than for a few years, as many couplings, particularly those involving the ocean, have a time constant of many years; also ice limits creep but slowly. A number of the parameters the climate modeller is interested in calculating as dependent variables are imposed by the general circulation modeller as boundary conditions, eg sea surface temperature.

There is no doubt that interaction between ocean and atmosphere is one of the most important processes controlling climate. The models are also used to investigate the response of climate to natural and man-made perturbations and the Office has already carried out experiments to assess the likely effects of changes in solar radiation, soil moisture, and, in particular, changes in atmospheric carbon dioxide (all these with a general circulation rather than a climate model).

The Office model was used to test Charney's theory of the Sahel drought and the conclusion was drawn that the feed-back mechanisms were far more complex than he had allowed for. This is a typical use for these models (Gilchrist, 1978).

9. Coda

In this lecture I have tried to indicate the wide range of scientific studies in the field of climate, how important observational studies are and the necessity for close interaction between observational studies and climate modelling.

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TROPIC OF CANCER
EQUATOR
TROPIC OF CAPRICORN
FIG 1 AVERAGE ANNUAL PRECIPITATION
CENTIMETERS
INCHES
Under 25
Under 10
25–50
10–20
50–100
20–40
100–150
40–60
150–200
60–80
Over 200
Over 80
Scale at 0 250 500 1000 1500 2000 2500 Miles
latitude 0 250 500 1000 2000 3000 4000 Kms.
FLAT P
QUARTIC EQ
PROJEC
ELEMENTS OF GEOGRAPHY by G. T. Trewartha,
and E. H. Hammond. © McGraw-Hill Book Co.

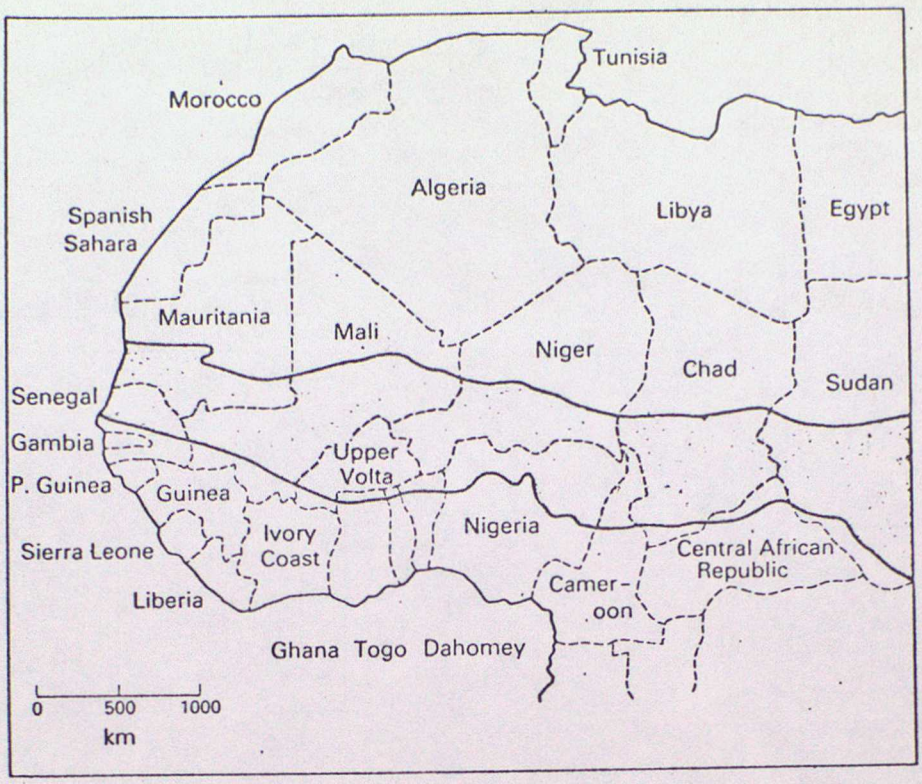

Figure 2 Sahelian zone countries.

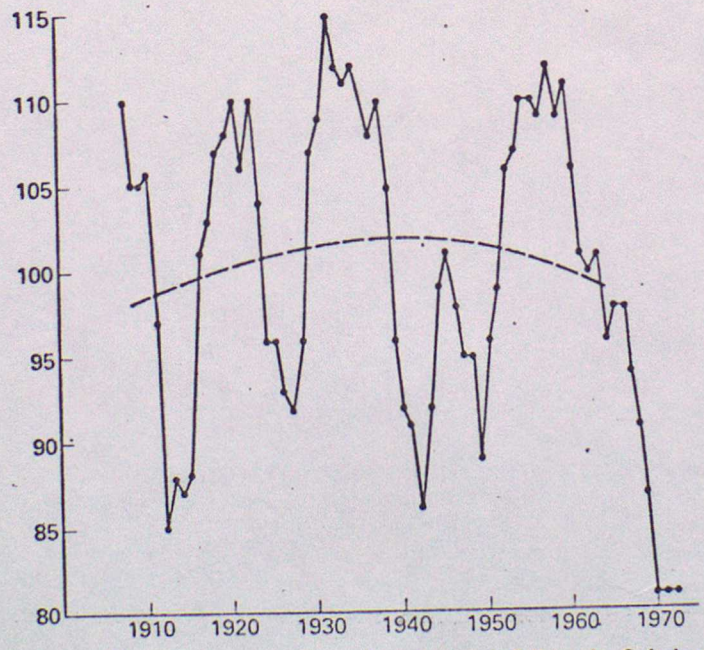

Figure 3a Five-year means of annual rainfall in the Sahel region of West Africa expressed as percentages of the 1905–70 average. The dashed curve shows the long-term trend (*after Mason, 1976*).

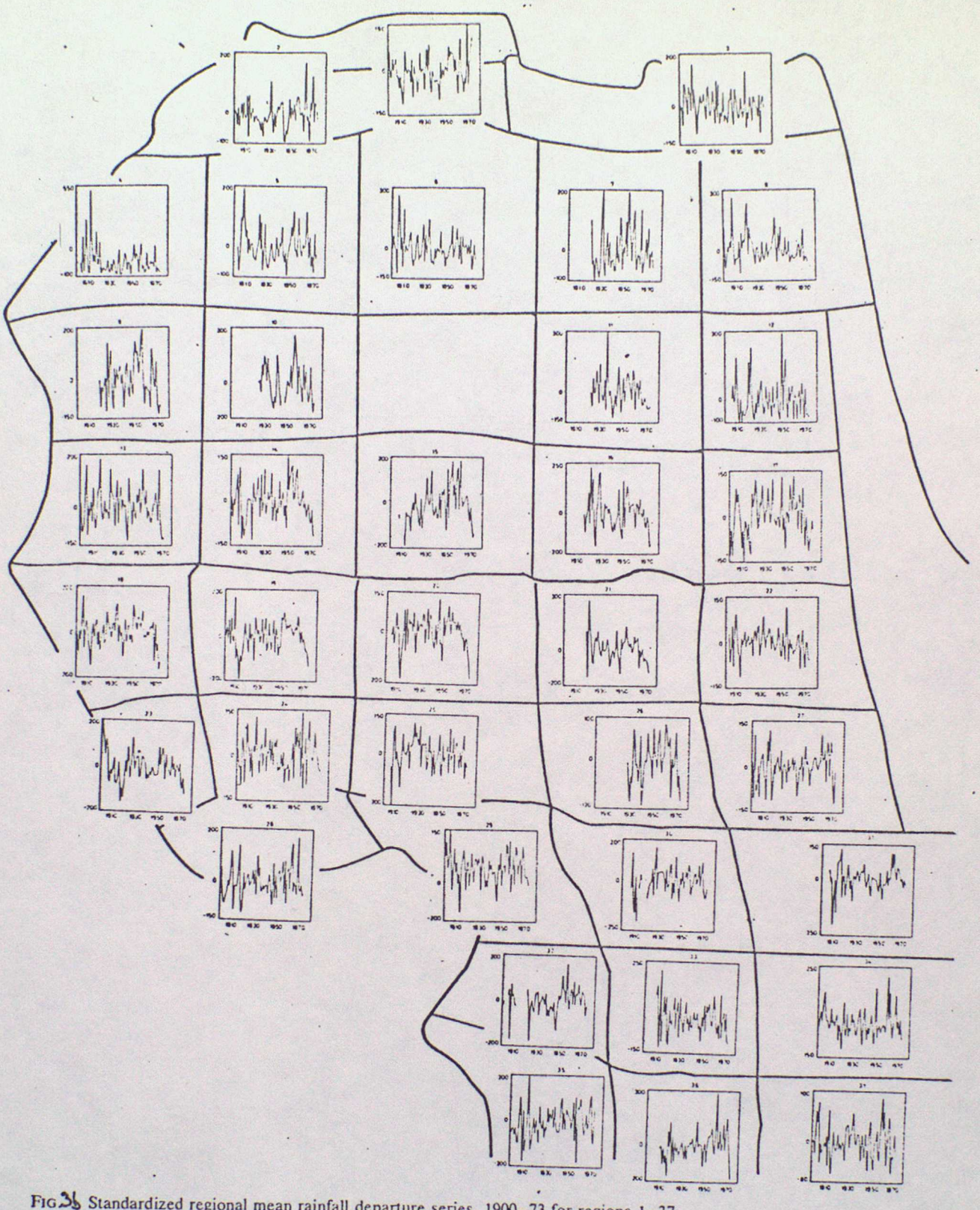

FIG 3b Standardized regional mean rainfall departure series, 1900–73 for regions 1–37,
Note distortion of longitudinal axis of outline map.)

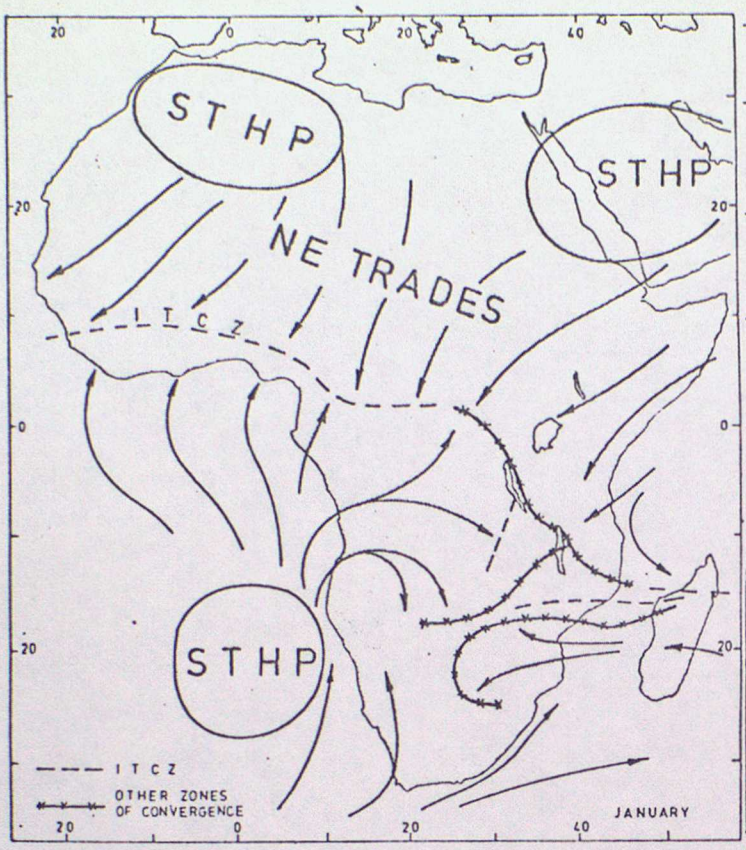

Figure 4a Features of the general circulation for Africa in January

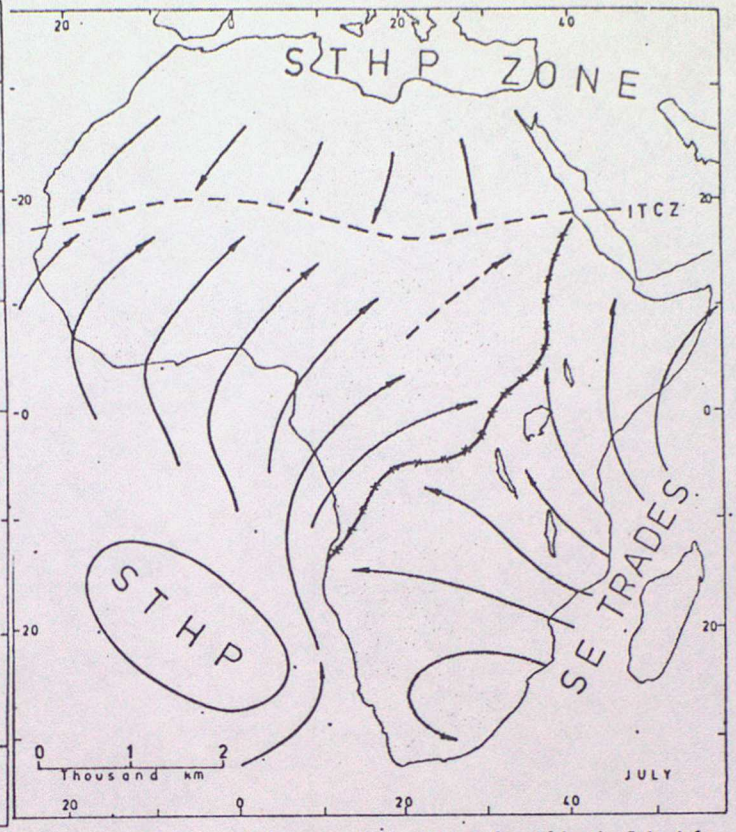

Figure 4b Features of the general circulation for Africa in July (after Hare, Griffiths *et al.*).

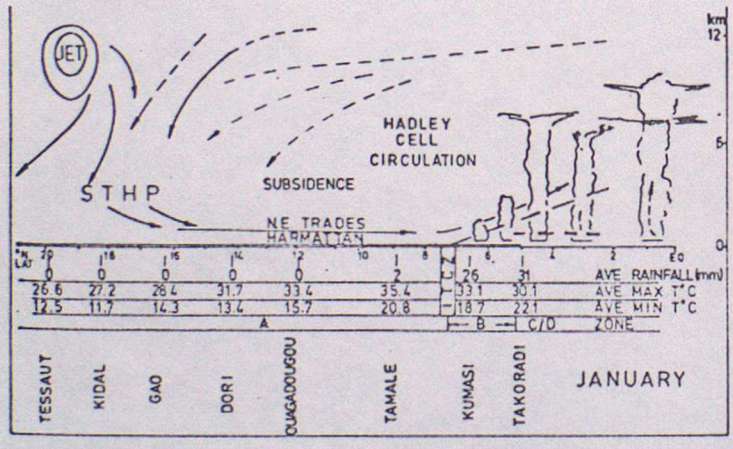

Figure 4c Meridional section through Central West Africa for January.

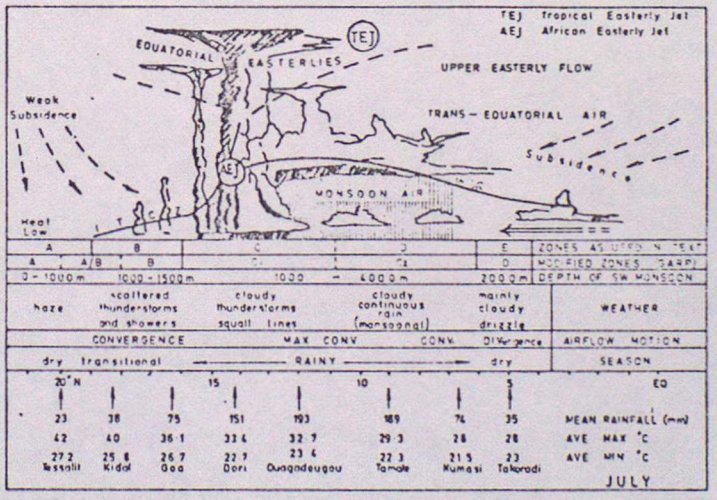

Figure 4d Meridional section through Central West Africa in July (partly after Dhonneur).

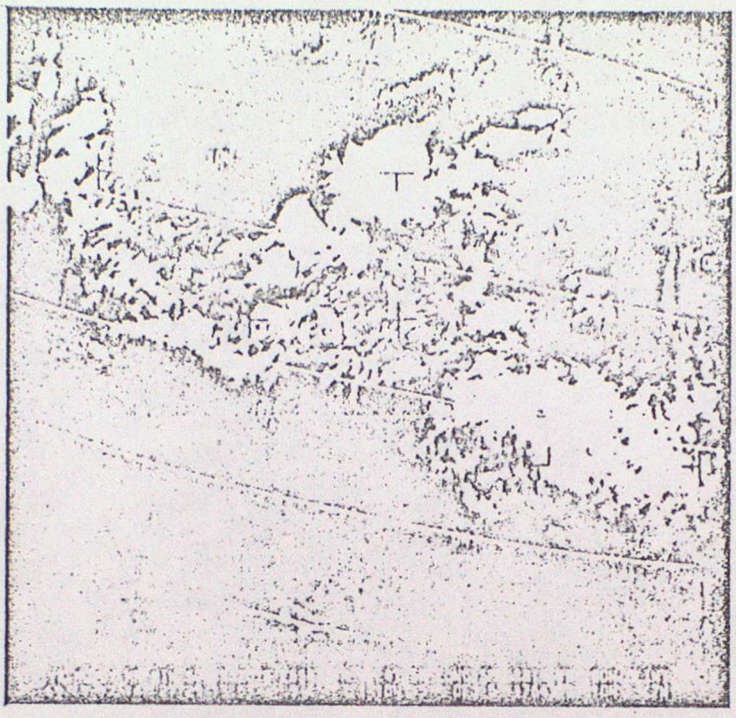

Plate 5 West Africa (South-West Monsoon). A varied cloud structure can be seen stretching inland from the south-west coast (centre left) of Liberia, Sierra Leone and Guinea. The main cloud band, vaguely striated south-west–north-east is about 300 km wide. Cloud is more broken over the Ghanaian and West Nigerian coast. Massed storm structures are evident at the head of the Gulf of Guinea (centre right), whilst cloud clusters dominate western Upper Volta (east of San) where rainfall in mid-July 1969 was higher than further south. Rainfall zones are not easily discernible from satellite data but a suggested zoning is indicated in the accompanying sketch map.

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ADVANCED LECTURES 1982 - THE CLIMATE OF THE WORLD

LECTURE 2 - Climates D. E. JONES

1. Introduction

Previous lectures have already indicated that the pictures of a simple symmetric general circulation pattern are far from the truth but do provide images and concepts from which to start discussion of the realities reflected in the observational aspects of the study of the general circulation. Figure 2.1 shows an idealised version of the surface wind field, Figure 2.2 an aspect of reality. We can recognise many correspondences, but also note for instance that the Intertropical Convergence Zone is north of the Equator in the east Pacific but a long way south in the southeast Indian Ocean. Likewise in the upper air the summer equatorial jet stream stretches from Southeast Asia to across Africa, but elsewhere the Equatorial easterlies are very light (Figure 2.3).

These asymmetries and irregularities, together with those introduced directly by the very varied topography of the land masses of the earth, by the surface temperatures of the circulating oceans make the climate at the earth's surface very complex and very interesting.

As Dr. Slingo has pointed out, when looking at Hadley cells in meridional cross sections of the atmosphere, one must remember that the zonal flow is very, very much larger so that an air parcel that remains in the Hadley cell is corkscrewing around the world; also the upward branch of the Hadley cell is really taking place in isolated cells of cumulo-nimbus clouds and other convective systems.

Features to be noted in figures 2.2 - 2.7 are:-

January surface pressure and wind fields - Subtropical highs; the low pressure trough of the Intertropic Convergence Zone near to or south of the Equator; the great anomaly of the Asian high with the surface air streaming away from it. This being a cold high, it decreases in intensity with height and is overlain by the midlatitude westerlies. In the southern hemisphere (summer) the subtropical highs form an almost continuous belt, with centres over the oceans, and heat lows dominate the three main land masses.

July surface pressure and wind fields - the northern subtropical highs are further north with considerable pressure differences between land and sea in the same latitudes: heat lows over southern United States, the Sahara and southern Asia. Figure 2.8 shows that the ITCZ moves further away from the Equator over land than over the oceans. Note that the lows decrease with height. Also note that the ITCZ, where the trade winds from each hemisphere meet, lies along troughs of low pressure; when the ITCZ is well away from the Equator the air which crosses the Equator curves from Easterlies to Westerlies but the curve takes place nearer to the Equator than to the pressure trough and the ITCZ (see figure 2.9 and The Indian Ocean in figures 2.2 and 2.4).

850 mb stream lines (note limited meridional extent of charts) the easterly Trade Winds are very evident in both seasons at this level except over the Indian Ocean.

200 mb stream lines (note limited meridional extent of chart) the westerlies dominate the winter hemispheres; in the southern summer the upper highs over the continental heat lows manage to generate some easterly flow near the Equator, but in the northern summer the flow of air out of the Indian monsoon becomes very intense and the part that flows southwards curves towards the east giving, between 250 and 150 mb, an elongated jet stream - the Tropical Easterly Jet. It is thought that the convergence at the right exit of this jet and the consequent subsidence help to make the Sahara dry zone more intense and more extensive.

In the time available I can do little more than outline the main climatic types and relate them to the General Circulation: I shall discuss some features of interest and point out some of the anomalies which require explanation.

2. Classification

The climate of even a small area is composed of a large number of climate elements and it is nearly impossible for two places to have identical climates so a condensation and ordering is needed to facilitate scientific analysis and explanation. Classification consists of recognising important characteristics which individuals have in common and grouping the individuals into a few classes or types in order to establi generalisations.

Classification can be based on the mode of generation of climates or on their effects. The genetic type appeals to meteorologists but it has to remain simple and the facts of climate may be forced into a preconceived genetic structure. Empirical classification starts from the observed facts but the basis of division is arbitrary and has to depend on the use to which it is to be put, and there is a risk that "description without reference to genesis or origin soon becomes dull and tiresome - features which unfortunately characterise much of the climatic literature on regions" (Trewartha 1968).

The best known empirical classification is that due to W Köppen and a considerable number of the types and subtypes coincide with certain broadscale features of the atmosphere thus highlighting the numerous and important deviations from a "standard" world pattern. Köppen's system is known throughout the world and was the result of half a century's study and refinements (see Trewartha 1968, Encyclopaedia Britannica, Lamb 1972). This classification is quantitive and is based upon annual and monthly means of temperature and precipitation; it accepts native vegetation as the best expression of the totality of a climate, so that many of the climate boundaries are selected with vegetation limits in mind. It recognises that plant development depends on available water, as well as temperature, so precipitation is modified by evaporation which is parametrized by temperature. Many other authors have proposed modified Köppen's schemes; they mostly differ in the method of expressing the available water.

Although I shall use a genetic approach in describing the various climates in that I shall relate them to the general circulation, the actual definitions will be based on an empirical classification due to Trewartha (1961/1981) based on Köppen's approach. (See Tables I and II for descriptions and Figure 2.10 for a chart).

Trewartha's scheme is a fairly typical one; 5 classes are defined in terms of monthly mean temperatures (basically freedom from any frost and length of the growing season) and one extra class which is defined on aridity and cuts across all the temperature classes except the polar class. (Dry climates are those where precipitation is less than potential evaporation, which is itself expressed in terms of monthly mean temperature - highly suspect but serves its purpose.) Like most schemes it sub-divides classes according to whether or not there is a dry season and when in the year it occurs.

Trewartha warns against drawing boundaries too precisely and shows an example of boundaries based on yearly averages (see Figure 2.11).

3. Tropical Climates

Let us first consider the air flowing around the oceanic cells of the sub-tropical high pressure belt centred about 25-35°N and S. The Equatorward-moving air on the eastern side and the air on the equatorward side of, and not too far from, the centre is stable and dry as it is under strong subsidence and a low inversion (Fig 2.12). Incidentally this gives a sunny and fairly dry climate to subtropical oceans and islands such as the Canaries, and it also causes the extremely arid coastal regions on the western sides of the continents in latitudes 20° to 30° N and S such as the Kalahari and Chile-Peruvian deserts and the western margin of the Sahara.

The Equatorward-moving air picks up heat and moisture from the sea, the inversion steadily rises and the air turns more and more to the west (Trade Winds) and becomes a deeper, unstable, wet airstream. As the airstream approaches the Equatorial trough of low pressure, and the corresponding flow from the other hemisphere, it is subjected to convergence and lifting and this leads to the great convective cloud systems of the Intertropic Convergence Zone (ITCZ or ITC). Thus although the Trade Winds are mostly dry, fair-weather winds on the east and poleward sides, their Equatorward and western oceanic sectors have many weather disturbances (even tropical storms in the western parts away from the Equator) and greater amounts of cloud and precipitation, culminating in the ITCZ itself. Note that much of the energy released in the upward branch of the Hadley cell has entered the atmosphere away from the Equator and is conveyed to the ITCZ by the Trade Winds. Also the averaged upward motion is concentrated in the convective systems and does not come about by a uniform gradually rising circulation.

The ITCZ follows the sun on its annual path, moving much further in some longitudes than in others, so a fundamental division of tropical humid climates is between those regions near the Equator which have considerable rainfall all the year round (Tropical wet climate or Equatorial rain climate) and those which have a dry season in the winter when the ITCZ has moved to the other hemisphere and the region is under the subsidence for the subtropical anticyclone. (Tropical wet and dry). This difference has a marked influence on the vegetation making the difference between the Jungles, hardwood forests and swamps of the Congo, the Amazon Basin, Southeast Asia and parts of Central America and the lighter, more deciduous forest and tree studded grasslands, often called Savanna typically located in South Central Brazil, northernmost South America and tropical Africa. Note it is not known whether the lack of trees in the grasslands in this climate is caused by the drought conditions in winter or by the frequent fires that sweep the grasslands from which the trees do not recover (Figure 2.8)

You will find that much modern work on the General Circulation and climate is concentrated on the Tropical and Equatorial zones: this is the major 'driving' area for the atmosphere - it covers a wide area - it is the zone about which least is known and, although it was once thought that tropical climates were simple and followed regular patterns, this is now known not to be true; above all the climates are not easily explained.

Africa is about the best example of the expected arrangement of climatic zones in the tropics, so we will look at the structure of the atmosphere over West Africa.

(Figures 4a - 4d of Lecture 1 refer)

Note: a) Southeast Trades cross the Equator to become moist, south-westerlies (sometimes called 'monsoon' air).

b) The northeast Trade Winds over the Sahara which are very dry.

c) The ITCZ does not normally get south of the Equator in winter; this is because of the great heat of the Sahara Desert. This means that the Guinea coastal area remains wet in the winter (about 1 inch in January).

d) In the eastern part of Africa the interactions of the air masses are more complex and are much affected by the high land of the region and, in summer, by the Indian monsoon flow.

Walker (1958) and Dhonneur (1971) have proposed five zones corresponding to five recognisable types of weather:

Zone A - Dominated by Sahara air, usually free of cloud.
Zone B - Within the wedge of moist air, but this is not deep enough (3,000 - 5,000 ft) to produce showers. Winds at 3,000 ft are usually north-east (Trades).
Zone C - The Equatorial south-westerlies are 10,000 - 14,000 ft deep and contain widespread convection. There is convergence at many levels. There are squall lines and 'tornadoes' with violent storms. Three or four different air flows may be involved at different levels and analysis is difficult. Many days are free of showers, and insolation is high.
Zone D - Similar to C but the deep south-westerlies are overlain by drier southerlies: cloud and rain tend to be more continuous than in C.
Zone E - Dominated by subsiding air from the subtropical high pressure region of the Southern Hemisphere.

Note, in the cross section for the summer, the Tropical Easterly Jet stream and the much lower African Easterly Jet which is due to the intense thermal gradient caused by heating over the Sahara. What effect these jet streams have on the meteorology of these areas has not really been worked out.

Quite untypical of these latitudes is the dry zone of tropical East Africa, for about 10°S to over 10°North, and this has been the subject of much study. In short explanation, the deficiency south of the Equator is almost certainly a consequence of the anomalous flow caused by the Indian monsoon, which replaces the moist easterlies experienced by other continental eastern margins by Southwesterlies that do not penetrate inland. The lack of rain in the South Somalia desert, in the winter seems related to the air stream having a trajectory from the Northeast, because of a high over Arabia, instead of across the warm ocean, but the lack of rain in the summer does not as yet have an adequate explanation, although a cold sea may be part of the explanation (Figure 2·13).

Sometimes these Tropical wet and dry climates are known as monsoon climates, not because they are caused by direct heating of land masses as in the Indian monsoon but because there is a marked reversal in wind direction between the Trade Winds of the dry season and the so-called Equatorial westerlies which develop in the ITCZ when it is well away from the Equator. (See Figure 2·9). Some people would argue that the marked seasonal wind shift of the Indian subcontinent is in the genetic sense no more a monsoon than those of other Tropical wet and dry regions but more a latitudinal migration of planetary wind belts following the course of the sun.

Incidentally although much of India is here classed as Tropical Wet and Dry, the dry season on much of the west coast is short and there is enough soil moisture during the dry months to support tropical rain forest. However the true Tropical wet climate starts further east in Burma.

A typical example of Tropical Wet (Equatorial rain) climate is that of Belèm at the mouth of the Amazon. Mean temperature of every month is near 26°C (79°F) and average monthly rainfall varies between 94 to 436 mm (4 to 17 inches). Sunshine averages 5½-6 hours a day.

The desert islands near the Equator in the Central and Eastern Pacific represent one of the most extraordinary climates in the world. They owe their dryness to the fact that a cool sea is normally maintained by the ocean circulation and the ITCZ is almost always north of the Equator (as a consequence?) In occasional years the ITCZ forms, and remains, over the area for much of the high-sun season and the sea is warm giving intense convective activity: average yearly totals of 500-750 mm (20-30 inches) are quoted for many of the islands but in some years only 50-150 mm (2-6 inches) fall and in the few wet years there may be 2,500 mm (100 inches). This is one of the most interesting of all the problems of climatology and will be referred to in other lectures. (See Figures 2.14, 2.15).

The tropics are characterised not only by high temperatures throughout the year but also by an annual range of daily mean temperatures which is less than the diurnal range of temperature: the seasons are characterised as wet or dry; in middle latitudes temperature as well as rainfall determine climatic type, and the dormant season for plant growth is usually one of low temperature rather than drought.

4. Subtropical Climates

On the western margins of all the continents, North and South, there are regions which in the summer are dominated by the anticyclones of the desert zone, particularly the stable eastern ends of the subtropical oceanic anticyclones, but which suffer, in the winter, mid-latitude travelling cyclones and the consequent changeability. These are the areas of Mediterranean climate and they are characterised by (1) a concentration of the year's modest amount of rain in the winter season, while the summers are nearly or completely dry; (2) warm to hot summers and mild winters; and (3) abundant sunshine especially in summer. The five regions where it occurs are notably similar: the borderlands of the Mediterranean Sea, the central and coastal strip of California, central Chile, the southern tip of South Africa, and parts of southernmost Australia.

In California and Chile the Mediterranean climate is ended abruptly by mountains with mountain-induced desert on the other side. The remaining continents of the Southern Hemisphere only just reach the appropriate latitudes, while in the Old World the Mediterranean Sea itself allows the climate zone to stretch for 2,500 miles, on to the land area as it acts as a cool-season convergence zone and route of winter cyclones.

We expect Mediterranean climate summers to be hot but it should be realised that at some coasts, notably in California and Chile, the sea is very cool and summer temperatures at the coast may be 10°C (60F) lower than similarly sited places elsewhere, with marine fog and low stratus, and also very strong sea breezes because of the contrast with the hot interior. Interior California is true to type for this climate as are the North African and Asiatic sectors in the Mediterranean rather than the three great peninsulas.

At the other end of the continents the corresponding regions receive in the summer the unstable or neutral Tropical Maritime air at a position where the subsidence of the subtropical anticyclone is weak. This favours abundant summer rain especially from convective activity and this is helped by a tendency for monsoon circulations to develop especially in South East United States and Southeast China. This climate which is also represented in all three southern continents, differs from the Mediterranean by having Tropical humid climates on the equator side, instead of hot dry climates, providing a source of humid sultry heat in summer instead of parching dry heat, and on the poleward side a severe continental climate which can bring intense cold (for the subtropics) in winter instead of the mild westerlies. This is the antithesis of seasonal precipitation patterns in the Mediterranean climates.

Summers are distinctly hot, average hot month temperature 75-80°F (24-27°C) with high humidity from the dominating Tropical maritime air, and nighttime temperatures high. Summer is the dominating season. Annual rainfall is relatively abundant 30 to 65 inches, much in summer from thunderstorms and organised convective systems; also tropical cyclones occur in late summer and the autumn.

In Winter there are incursions of Tropical Maritime air associated with travelling depressions and, in the northern hemisphere regions, strong monsoonal outpourings of cold polar continental air which help give Shanghai, for instance, average January temperatures of 38°F (3°C) and occasional severe freezes to the Gulf States of the USA. The southern hemisphere regions do not experience these cold outbreaks.

The meteorology of places like Southeast Asia and the interaction of the various air masses is extremely complicated (Figures 2.16a & b); causes and effects are often not understood, and are often the subject of disagreement. In winter the subtropical branch of the jet stream flows around the southern slopes of the Himalayas, while the polar-front jet occurs to the north of the plateau of Tibet, and they both meet in an upper convergence zone over China. This upper convergence zone leads to subsidence over inner China and therefore to generally fine weather in winter. The two jet streams tend to reinforce each other over Japan, creating extremely high wind speeds which can reach 100 m/sec at the 200 mb level.

In summer, the subtropical jet stream flows entirely to the north of Tibet and the upper convergence over China has disappeared. (Watts 1969, Ye 1981, Staff Members, Academia Sinica 1958).

5. Mid-latitude (Temperate) Climates

As we move further north we reach the middle latitude climates (called "temperate" by Trewqrtha but this word hardly applies to the continental subdivision where temperatures can be pretty extreme). These are dominated by the westerlies. The natural subdivision is into maritime (oceanic) climate, which of course is found at the western sides of continents, or the southern maritime extremities in the case of Australasia, and continental climate which is found only in the northern hemisphere, either well inland or at the eastern sides of the continents where severe seasonal temperatures result from their leeward location and the monsoonal wind systems.

The temperate oceanic climates lie poleward of 40° and stretch well towards the poles because of warm ocean currents; the depth of penetration into a continent depends on the terrain: the regions are very narrow in north and south America and in Scandinavia.

The major characteristics, of course, are cool summers, mild winters, changeable weather, considerable cloudiness and adequate rainfall at all seasons; that is not to say that all the months have the same amount but there is no seasonal emphasis and seasonal rainfall deficiency does not cause a dormant period for plant growth.

Points for comment are the unsually dry summers in the north west United States, probably caused by the unusually northerly extension of the North Pacific high; the exceptionally mild winters over large areas of Europe caused by the openness of the continent to Atlantic air and the warmth of the eastern North Atlantic: the rapid decrease with altitude of the length of the growing season (the Pennines at 1,700 ft resemble tundra). The Chilean temperate oceanic climate is a mountainous, cloudy, rain drenched region swept by strong winds.

In temperate continental climates (here defined as having a cold month under 32°F) a genuine winter, with a durable snow mantle, is combined with an authentic summer season, the annual range of temperature is large, with particularly low winter temperature. In the maritime climate one winter temperature is much like another, but wide deviations from the normal seasonal temperature are characteristic of continental climates-as much as 30°F in extreme climates. Summer is the season of maximum rain, which is mostly convective.

The variations on this class are so numerous that one tends to break into detailed description and one begins to notice the arbitrariness of classification. Seasonal temperatures and precipitation can vary considerably from year to year in the continental class, and even with the maritime version the positions of the dominating troughs and ridges in the westerlies, and the incidence of 'blocking', make a difference to the general character of a season. These climates are really only describable in terms of Synoptic Climatology i.e. the incidence of various synoptic features and their associated weather types. A few features of the North American continent will be taken as examples.

We note first that whereas in Eurasia moist oceanic air can penetrate from the west with comparative ease, and the Mediterranean and the Black Sea provide water vapour (and heat in winter) which extends the maritime influence well into the continent, in North America the north-south alignment of the Rocky Mountains in the west prevents the transfer of unmodified Pacific air into the interior. On the other hand, the penetration of moist air from the Atlantic Ocean and Gulf of Mexico into the eastern United States is easy.

In winter both continents have a cold anticyclone as a prominent feature, both situated to the west of the climatological upper troughs (Figure 2.17). The anticyclone over Asia is firmly sited well away from the western seaboard and is undisturbed by any midlatitude cyclones. On the other hand the North American winter high tends to be sited over, or to the east of, the Rockies, which to some extent determine its position; it is not as stable a feature as the Asiatic high as it is often interrupted by moving depressions (Figure 2.18). However on those occasions when it is a reality, especially when it builds from a cold, moving anticyclone over northwest Canada, it brings intensely cold polar outbreaks to much of the eastern half of the continent (see Figure 2.19) just as surges in the Asian northeast monsoon bring intense cold to Japan and south China.

Figure 2·18 shows that winter depressions can take many tracks across the American continent. Numerous depressions originate on the eastern side of the Rockies: some of them are lee depressions, that is they form as disturbances in the Pacific, lose their identity over the mountains, and re-form under the travelling upper trough. Many of these lows track towards the area of the Great Lakes, which is a region frequented by depressions tracking from many directions. Another breeding ground for travelling cyclones is the Atlantic coast, near Virginia. Any of these depressions, if they get very deep, will give very cold weather over wide areas of North America.

In the summer, the circulation is weak and depressions are confined further north in Canada, although fronts may penetrate south across the mid-states of the USA.

6. Dry Climates.

The essential feature of a dry climate is that the annual water loss through evapo-transpiration at the earth's land surface exceeds the annual gain from precipitation: thus dry climates do not have a constantly replenished ground-water. We see from Figure 2.10 that the arid climates cover a broad latitude band.

Typically the deserts are surrounded by semi-arid lands (called steppe around the mid-latitude deserts in southern Russia), and here the rainfall is very unreliable, e.g. the Sahel.

The typical types of desert climates are:-

The tropical dry climates lying in the vicinity of 20° or 25° N and S with their north and south margins at about 15° and 35°+, so they coincide fairly closely with the subsiding air masses of the subtropical anticyclones with their dry outward-flowing winds. They do not reach the eastern coasts of the continents because of the onshore winds. There are two types of semi-arid regions around these deserts - those that lie on the Equator side and experience rainfall at the time of high sun from the ITC, and those which lie on the poleward side and receive rain from the mid-latitude westerly belt in winter. The semi-arid lands can receive up to 20 or 30 inches of rain in these latitudes. Typical desert rainfalls for this type are less than 5 inches (Cairo 1.2 inches; Yuma, Arizona, 3.3 inches; Calama, northern Chile, zero, apparently). Typical average summer month temperatures are 95°F (35°C), winter month, 55°F (15°C).

The cool western litorals, already mentioned, where cool ocean currents prevent convection. Typical average monthly temperatures about 10°C lower than normal for the latitude, with small annual range.

Mid-latitude deserts and steppes which are either well away from the sources of water vapour (the oceans) or sheltered from them by mountains lying across the prevailing wind flow. In the United States, arid climates approach relatively closely to the sea. In the Southern Hemisphere, only Patagonia (Argentina) reaches high enough latitudes and has high enough mountains to experience this climate. Representative annual rainfall for these deserts is 3 to 6 inches, and up to 15 inches for the steppes (this is less than the subtropical deserts because of the lower temperatures). Their temperature ranges are similar to those at other places in the same latitude.

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TABLE I

The 6 great climatic groups and their poleward boundaries[7]

Based on temperature criteria

A. Tropical: Frost limit in continental locations; in marine areas 65° (18° C) for the coolest month (after Wissmann)
C. Subtropical: 8 months 50° (10° C) or above[8]
D. Temperate: 4 months 50° or above
E. Boreal: 1 (warmest) month 50° or above
F. Polar: All months below 50°

Based on precipitation criteria

B. Dry: Outer limits, where potential evaporation equals precipitation

TABLE II

Scheme of climatic groups and climatic types

Groups of climate	Types of climate	Pressure system and wind belt		Precipitation
		Summer	Winter	
A. Tropical humid	*Ar*, tropical wet	ITC, doldrums, equatorial westerlies	ITC, doldrums, equatorial westerlies	Not over two dry months
	Aw, tropical wet-and-dry	ITC, doldrums, equatorial westerlies	Drier trades	High-sun wet (zenithal rains), low-sun dry
C. Subtropical	*Cs*, subtropical dry-summer	Subtropical high (stable east side)	Westerlies	Summer drought, winter rain
	Cf, subtropical humid	Subtropical high (unstable west side)	Westerlies	Rain in all seasons
D. Temperate	*Do*, oceanic	Westerlies	Westerlies	Rain in all seasons
	Dc, continental	Westerlies	Westerlies and winter anticyclone	Rain in all seasons, accent on summer; winter snow cover
E. Boreal	*E*, boreal	Westerlies	Winter anticyclone and polar winds	Meager precipitation throughout year
F. Polar	*Ft*, tundra	Polar winds	Polar winds	Meager precipitation throughout year
	Fi, icecap	Polar winds	Polar winds	Meager precipitation throughout year
B. Dry	*BS*, semiarid (steppe)			
	BSh (hot), tropical-subtropical	Subtropical high and dry trades	Subtropical high and dry trades	Short moist season
	BSk (cold), temperate-boreal		Continental winter anticyclone	Meager rainfall, most in summer
	BW, arid (desert)			
	BWh (hot), tropical-subtropical	Subtropical high and dry trades	Subtropical high and dry trades	Constantly dry
	BWk (cold), temperate-boreal		Continental winter anticyclone	Constantly dry

Undifferentiated highlands

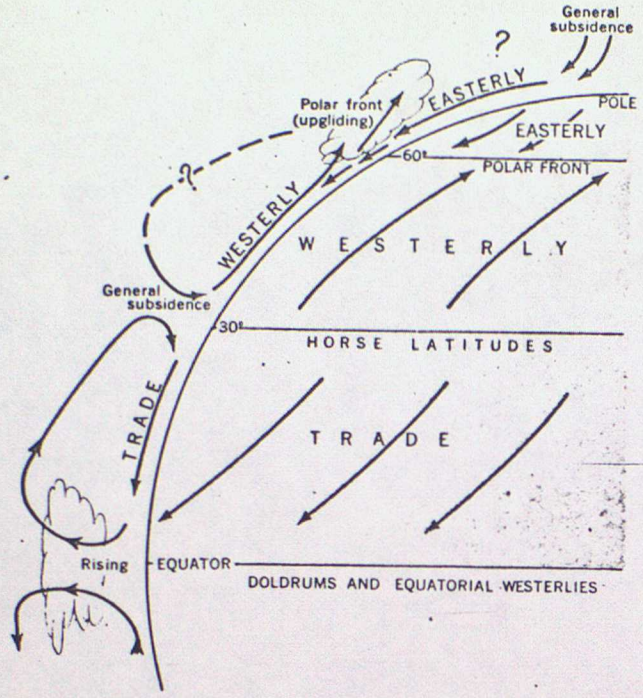

Figure 2.0 Important elements of the general circulation of the atmosphere. Some features of the upper circulation are not at present sufficiently well understood, or are too complicated, to be represented in a simple diagram.

Figure 2.1 A much idealized representation of the earth's surface winds. Average airflow is predominantly from an easterly direction in most of the low latitudes, and from a westerly direction in the middle latitudes.

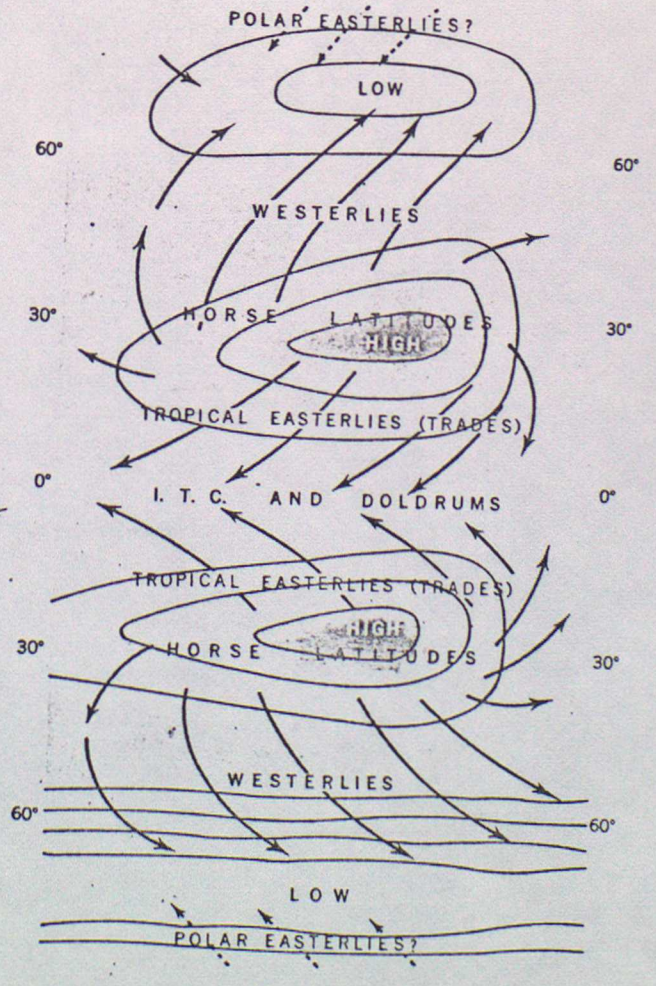

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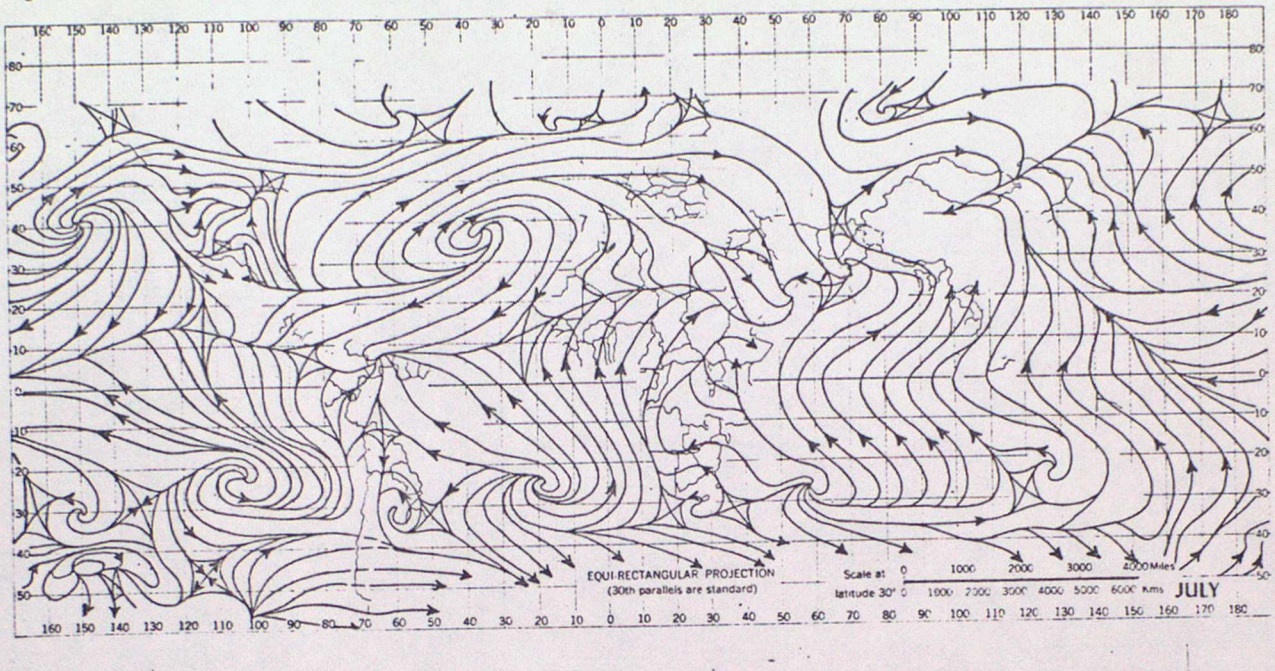
Figure 2·2 July: Direction of mean surface resultant wind. (*After Mintz and Dean.*)



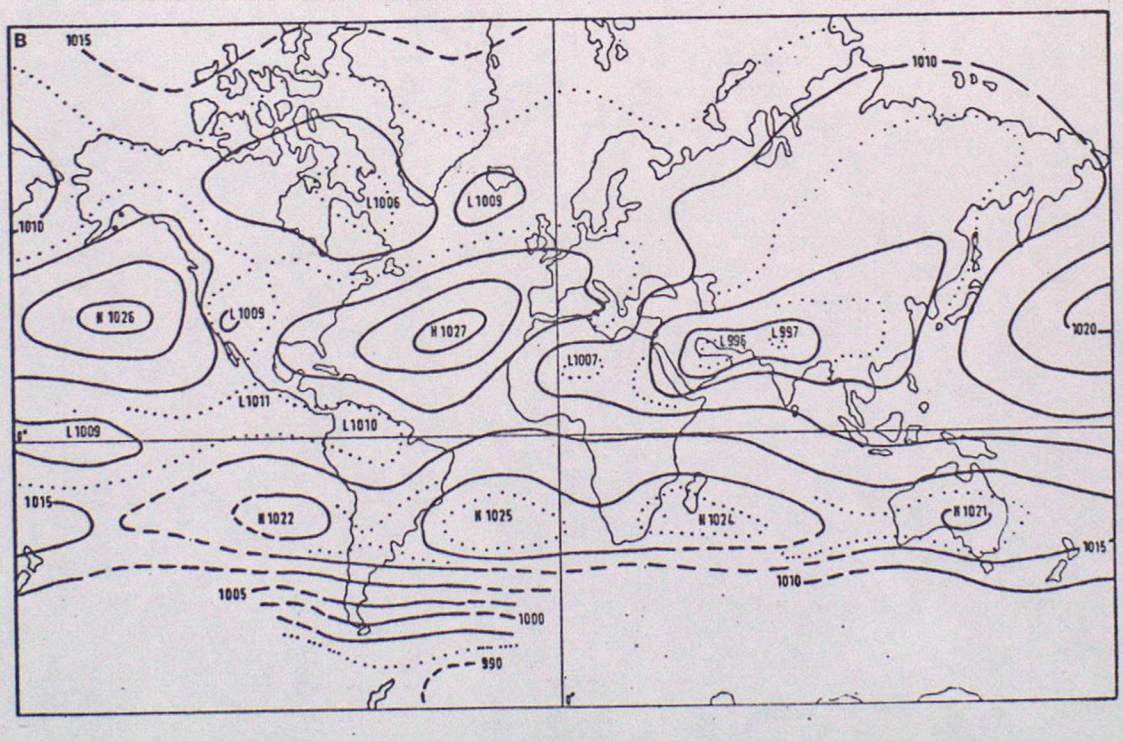


Fig 2·3 B : Average atmospheric pressure (mb) at sea level in July (1900–1939). (*After Lamb, 1969.*)

Fig 2.4a STREAMLINES 850mb JUN – AUG

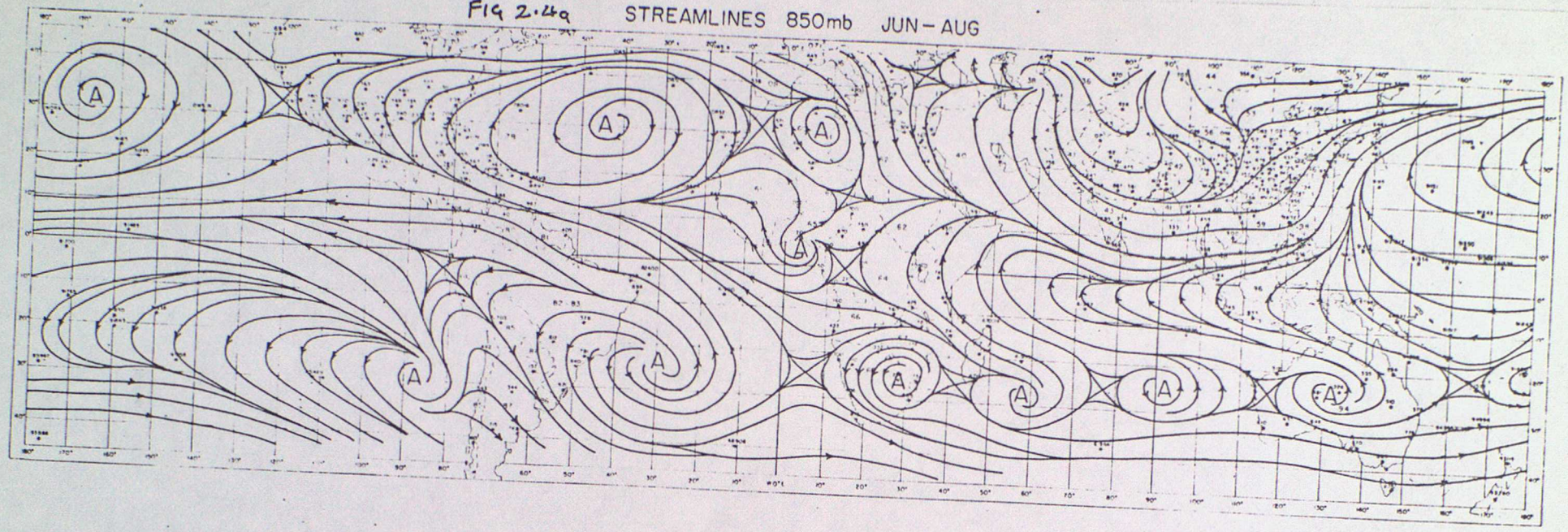
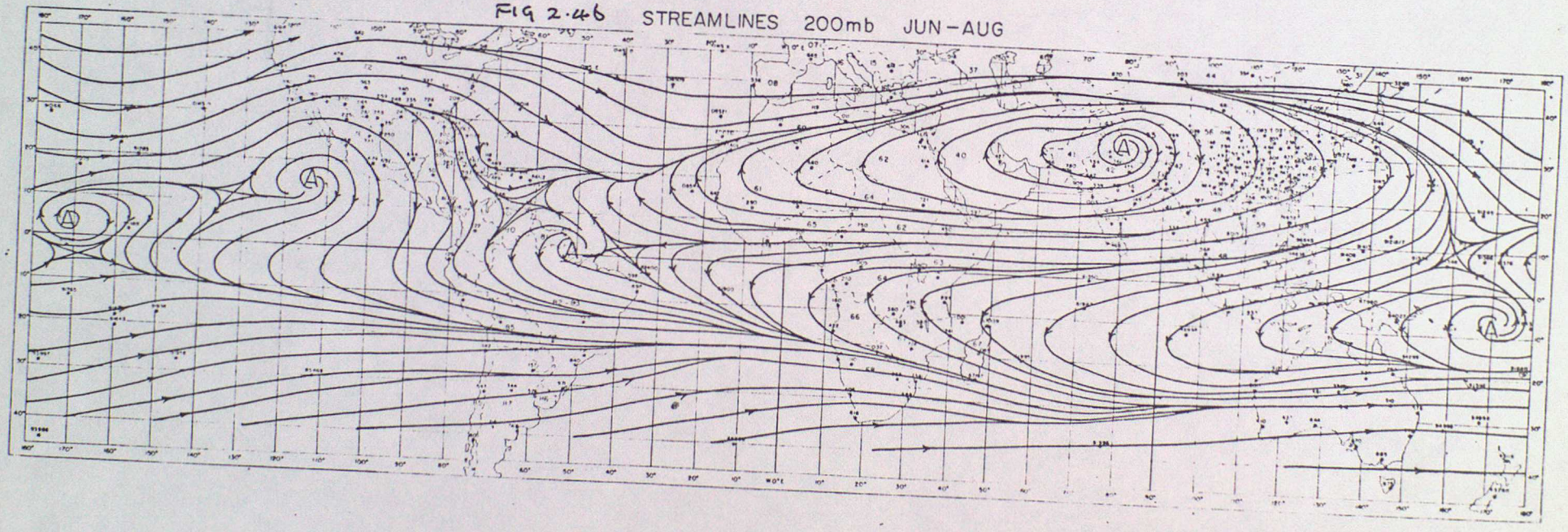


Fig 2.4b STREAMLINES 200mb JUN – AUG



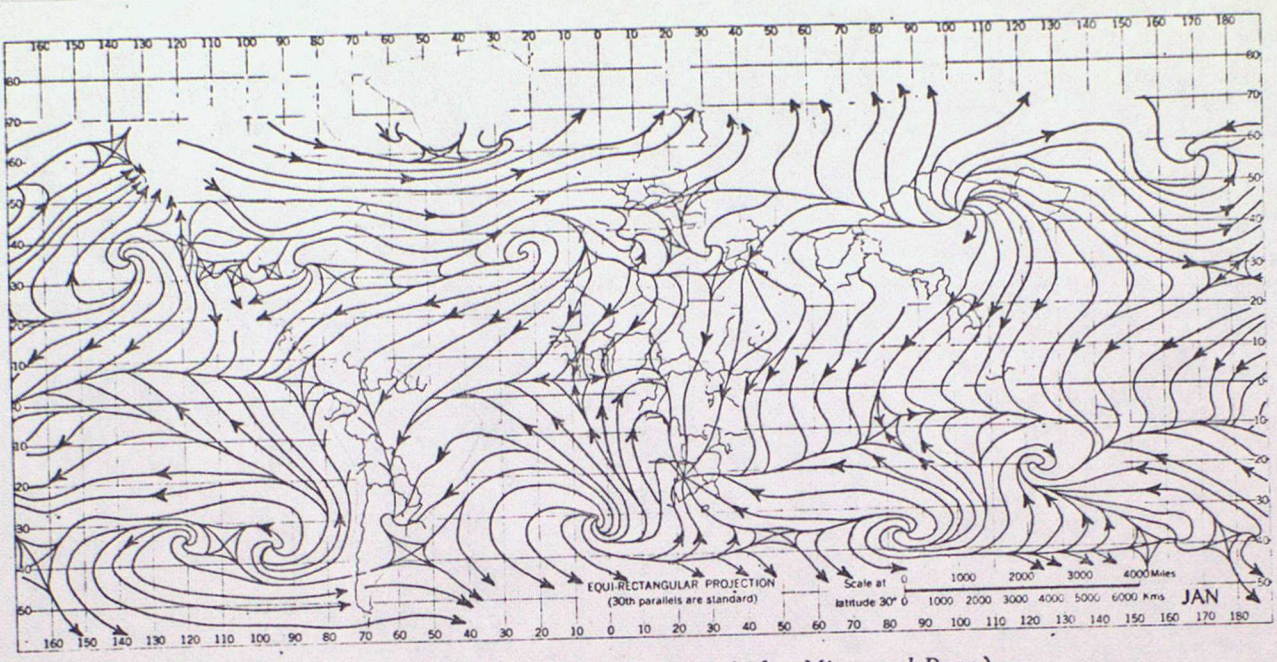


Figure 2·5 January: Direction of mean surface resultant wind. (*After Mintz and Dean.*)

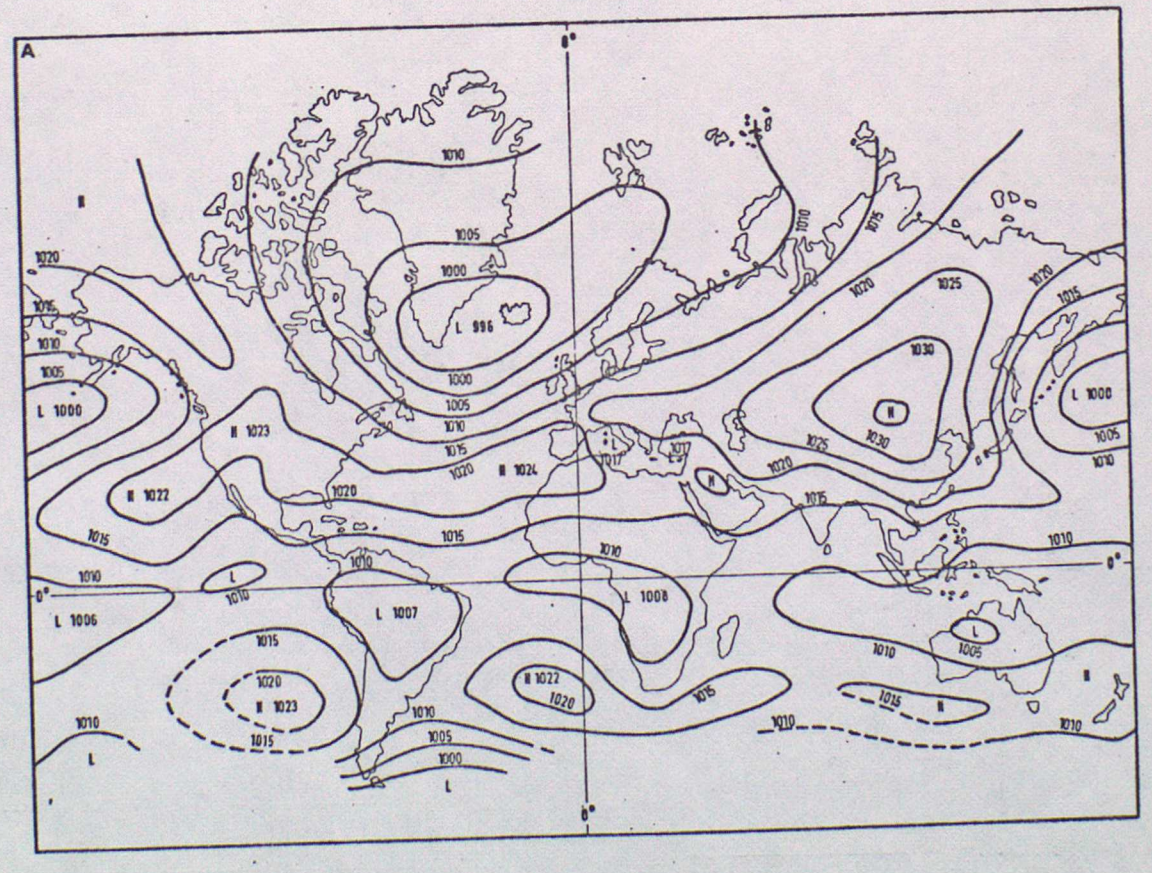


Fig 2.6 A: Average atmospheric pressure (mb) at sea level in January (1900–1939).

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FIG 2.7a STREAMLINES 850mb DEC-FEB

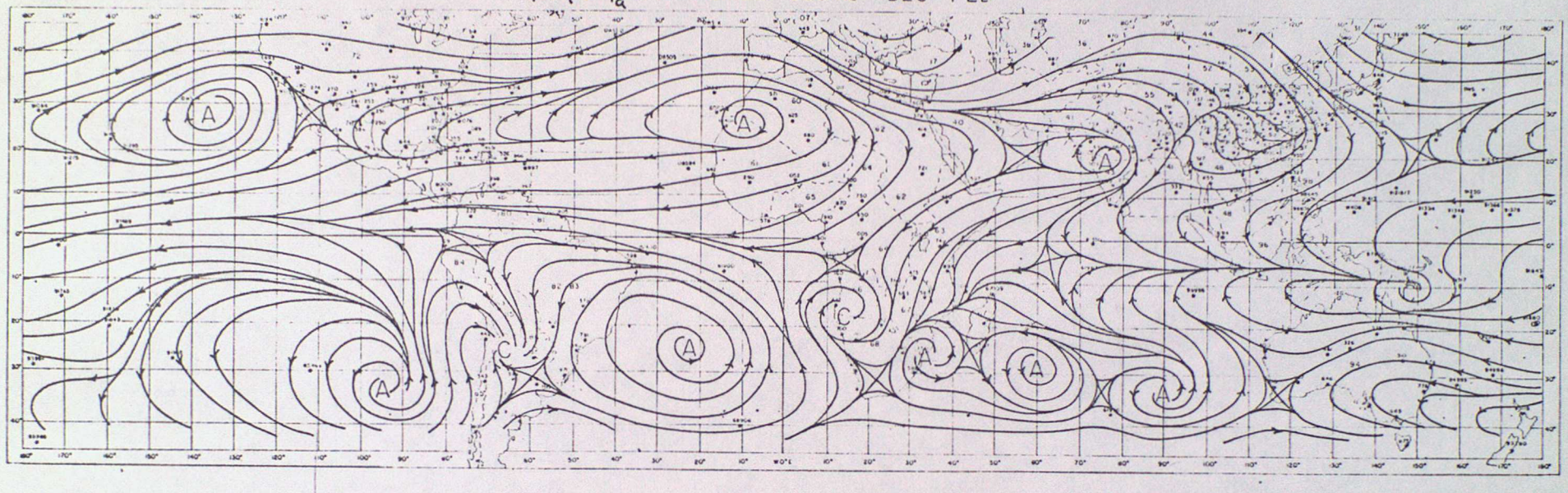
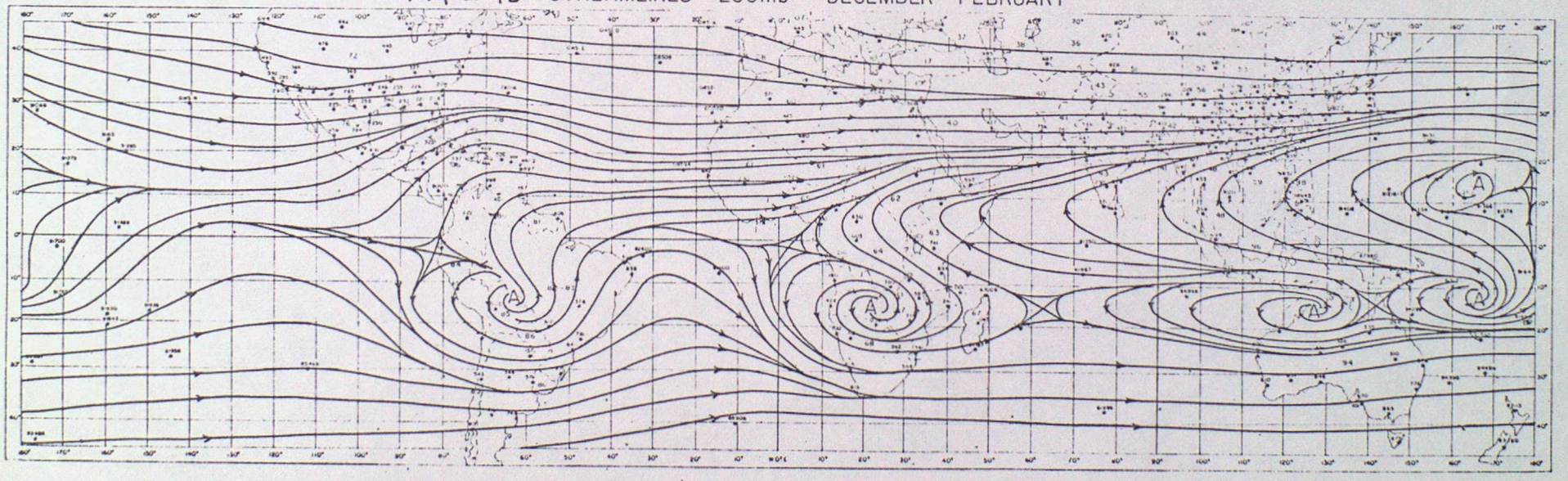


FIG 2.7b STREAMLINES 200mb – DECEMBER-FEBRUARY



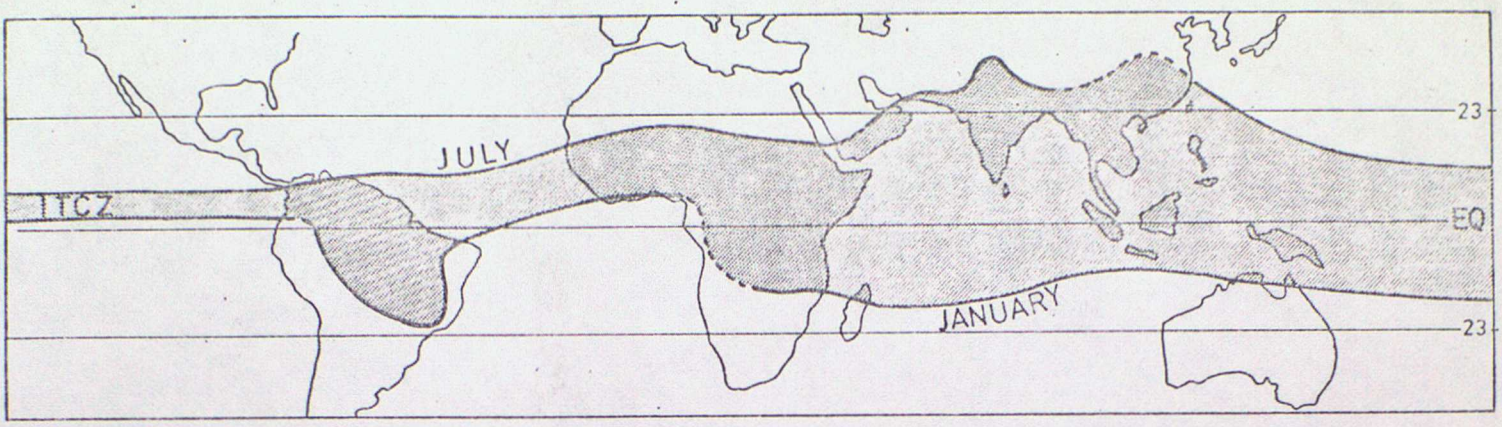


Figure 2.8 Mean position of the ITCZ in January and July (after Flohn).

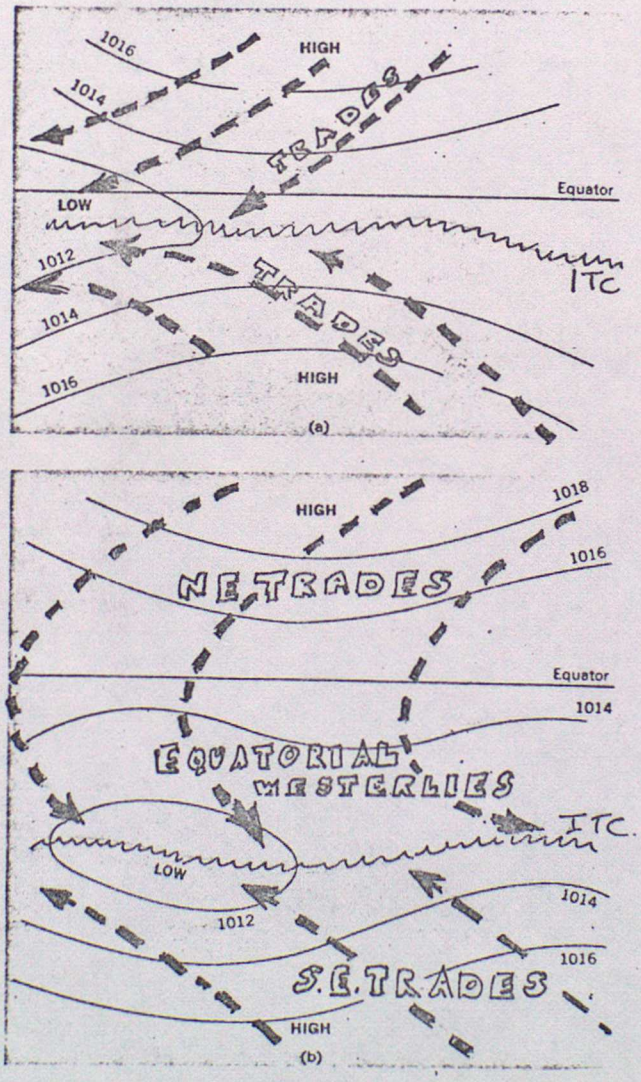


Figure 2.9 Pressure and circulation patterns *a* when the ITC is located near the equator, and *b* when the ITC is displaced some distance north-south of the equator. (*After Sawyer.*)

FIGURE 2·10
GROUPS AND TYPES OF CLIMATE
GROUPS
TYPES
A TROPICAL HUMID CLIMATES
Af Tropical Wet
Aw Tropical Wet-and-Dry
w, winter dry
s, summer dry
B DRY CLIMATES
BW Desert or Arid
BS Steppe or Semiarid
C SUBTROPICAL CLIMATES
Cs Subtropical Dry Summer
Cf Subtropical Humid
D TEMPERATE CLIMATES
Do Temperate Oceanic
Dc Temperate Continental
E BOREAL CLIMATE
E Boreal
F POLAR CLIMATES
Ft Tundra
Fi Ice Cap
H HIGHLAND CLIMATES
H
Extensive Tropical Uplands in A Climates Over 2000 ft. (610 m.) in Elevation
Scale at 0 250 500 1000 1500 2000 2500 Miles
0 250 500 1000 2000 3000
TROPIC OF CANCER
EQUATOR
TROPIC OF CAPRICORN
(Ice)
(Ars)
FLAT POLAR
EQUAL AREA P
ADAPTED BY ARTHU
ELEMENTS OF GEOGRAPHY by G. T. Trewa
and E. H. Hammond. © McGraw-Hill Boo
TREWARTHA, G.T.

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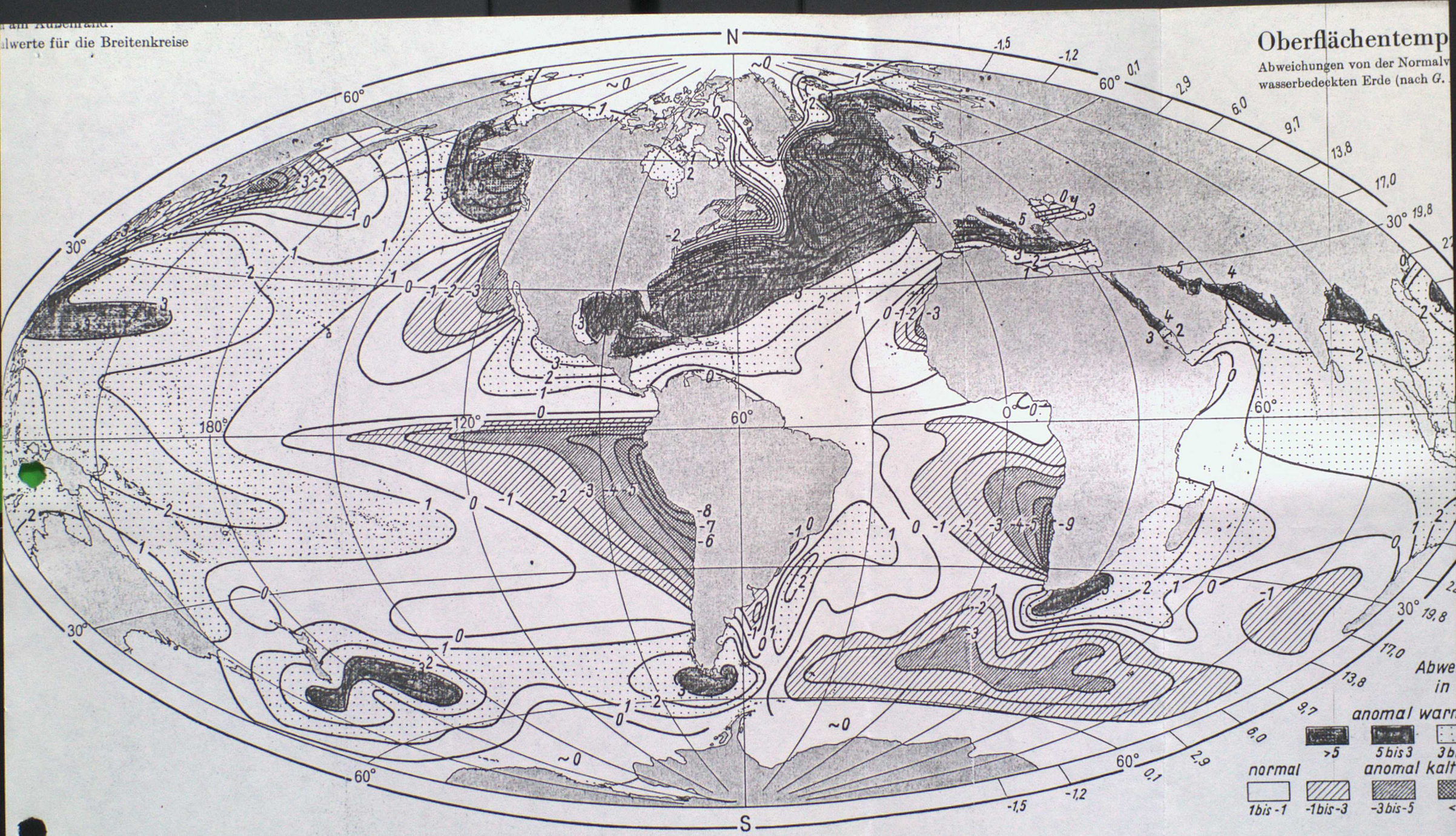


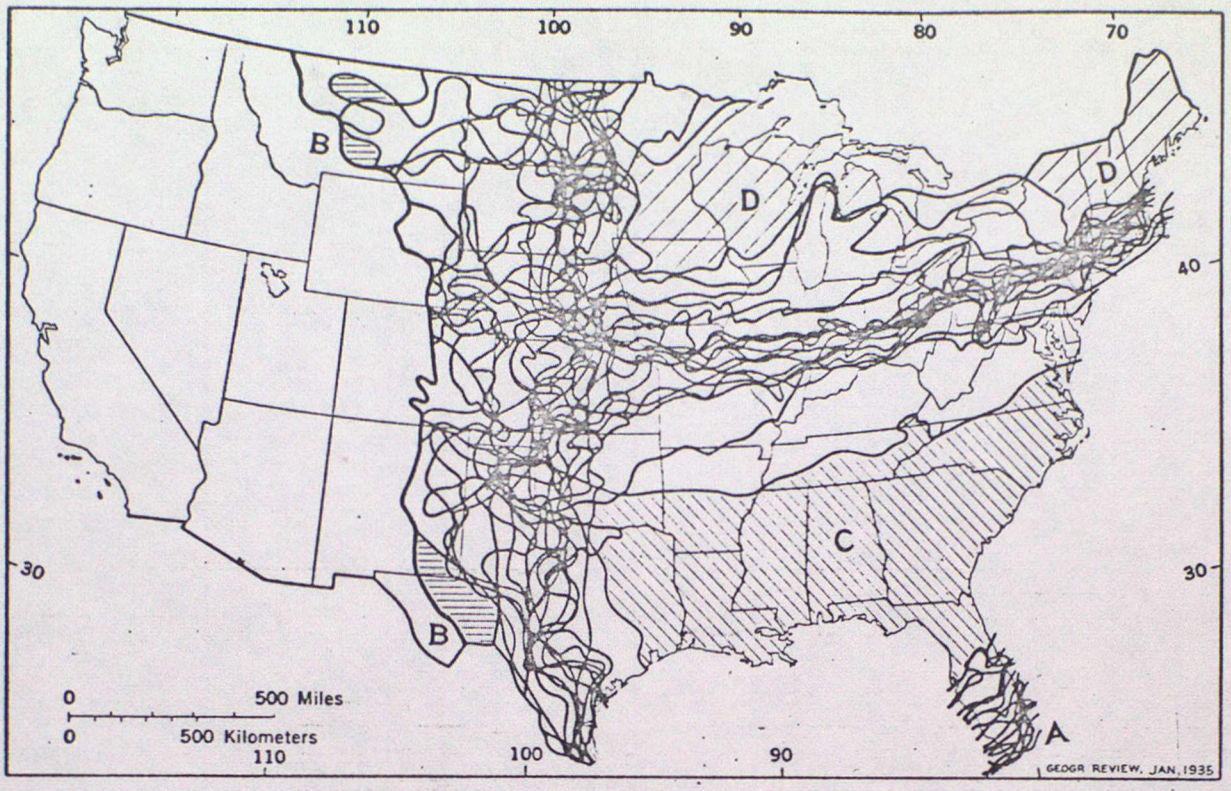


Figure 2.11 Climatic-year boundaries (Köppen system of classification) in the central and eastern United States. Each heavy line represents the climatic boundary between Köppen's dry (*B*) and humid climates, or between his mesothermal (*C*) and microthermal (*D*) climates, for a single year. Clearly, climatic boundaries shift position from one year to the next. Therefore the boundaries drawn on a map showing climatic types and regions (such as the map inside the front cover) are simply the means of different locations for a large number of individual climatic years.

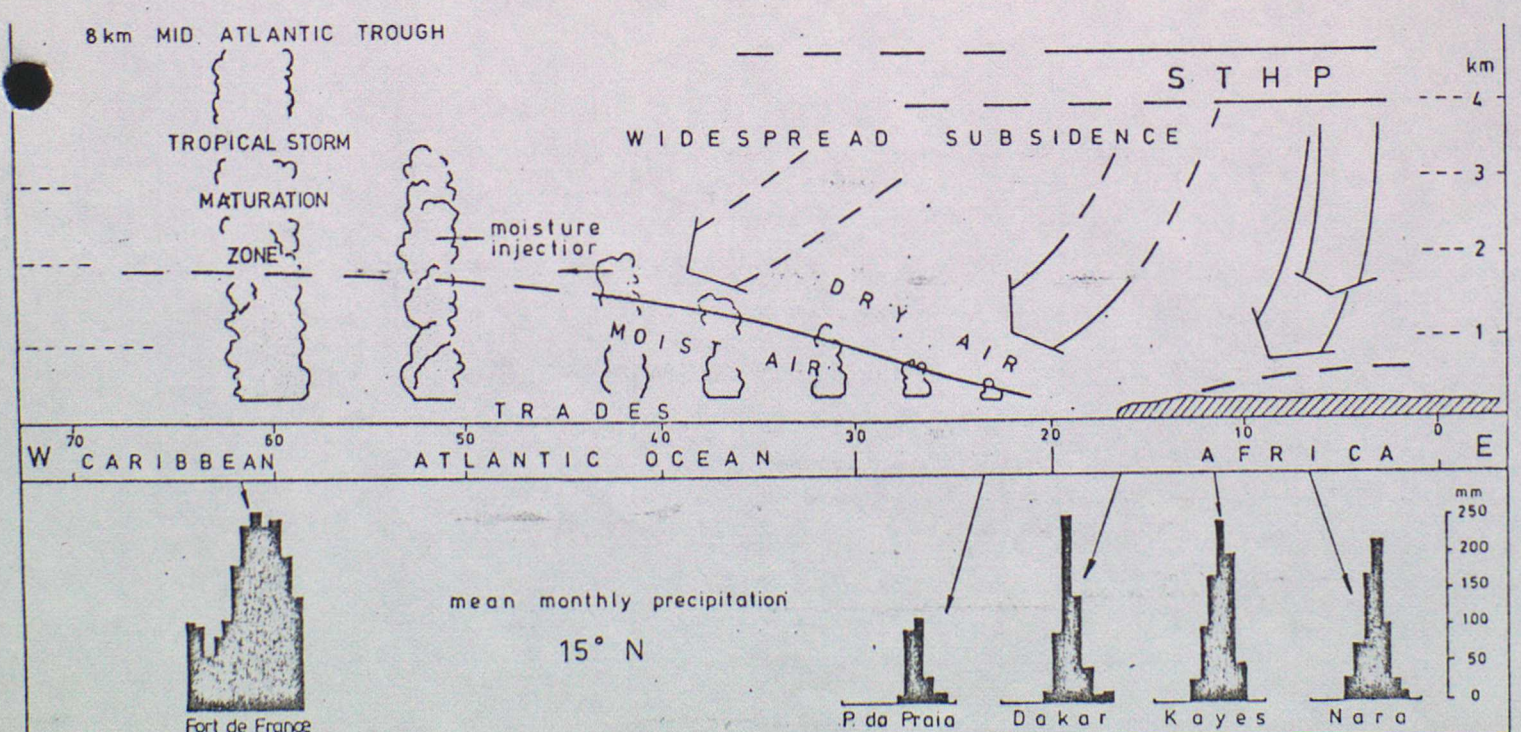
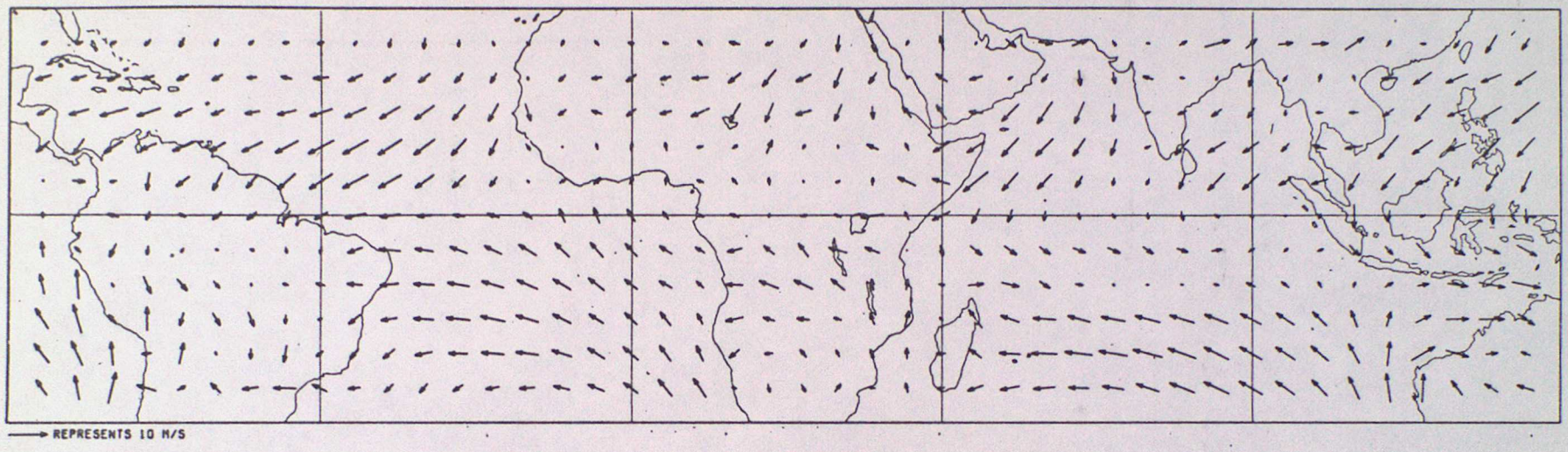


Figure 2.12 Atlantic cross-section (15° N) showing the structure of the low-level air flow in early summer. Rainfall graphs are given for selected stations.

MET OFFICE FGGE IIIA ANALYSIS
LOW LEVEL JANUARY MEAN WINDS



MET OFFICE FGGE IIIA ANALYSIS
LOW LEVEL JULY MEAN WINDS

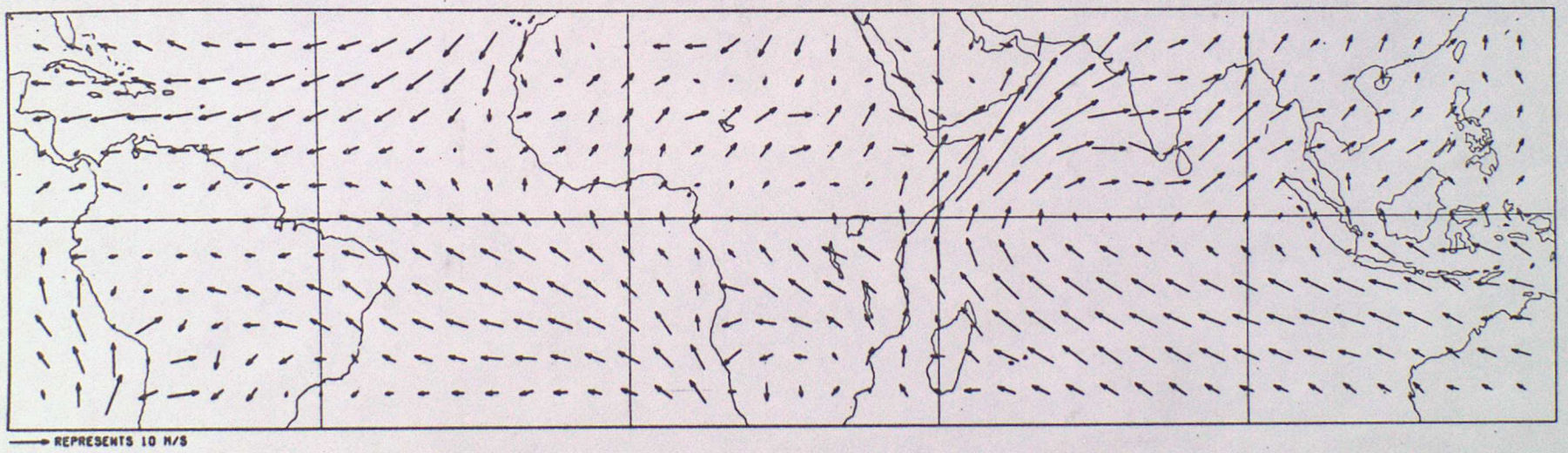


Figure 2.13 Winds at σ = .987 during 1979

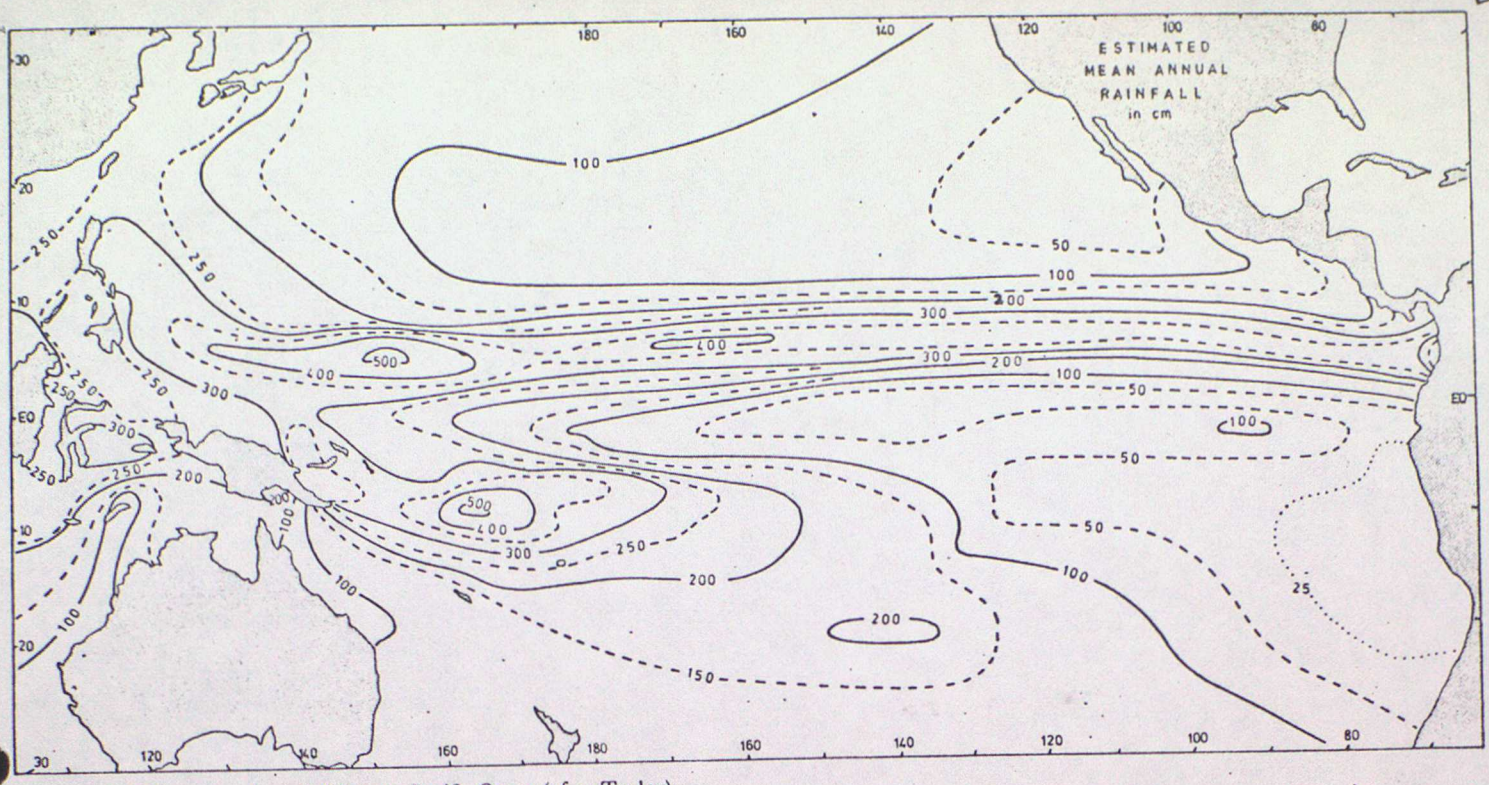


Figure 2·11나 Mean annual rainfall for the Pacific Ocean (after Taylor).

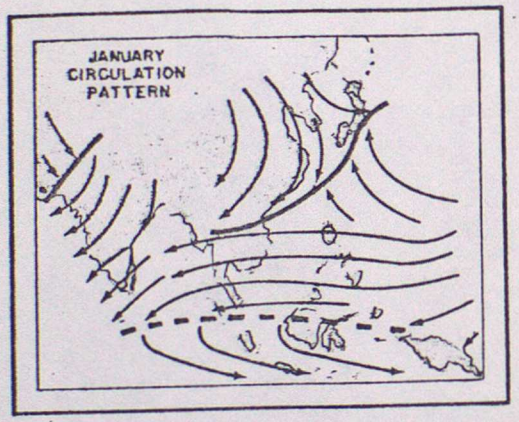


Fig.2·16a Principal elements of the low-level circulation pattern of eastern and southern Asia in the cool season. (After Thompson, Watts, Flohn, and others.)

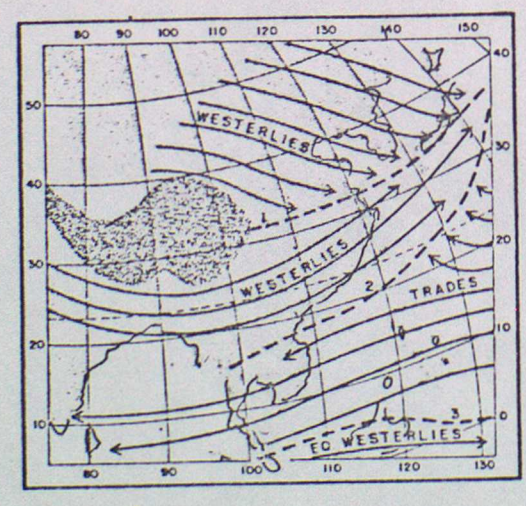


Fig2·16b Characteristic features of air flow at about 10,000 feet, November to March. 1. Tibet lee-convergence zone; 2. polar front; 3. *ITC*. (After Thompson.)

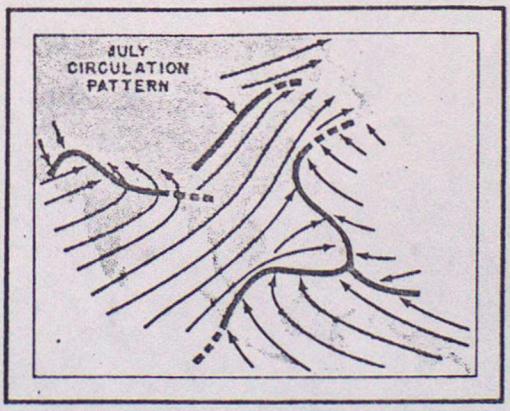


Fig.2·16c Principal elements of the low-level circulation pattern of eastern and southern Asia in the warm season. (After Thompson, Watts, Flohn, and others.)

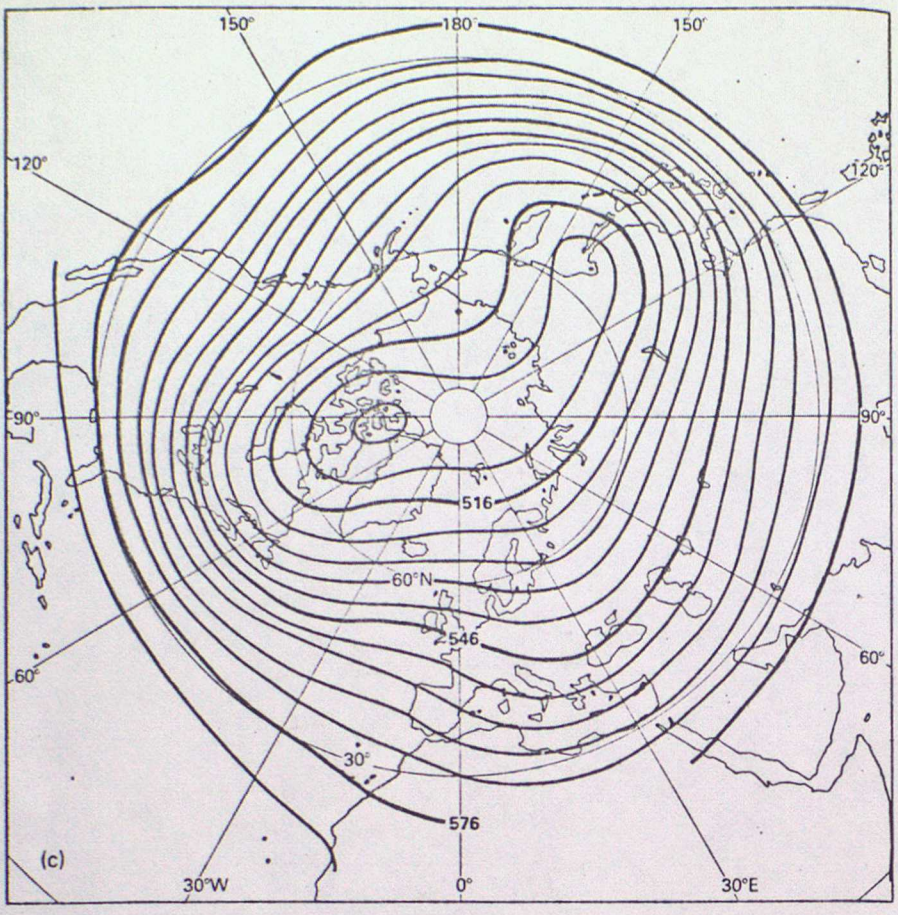


Figure 2.17 Monthly mean 500 mbar contours for January 1951–66.

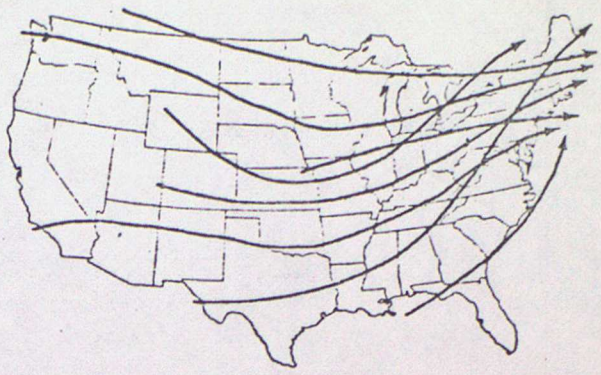


Figure 2.18 Main tracks of cyclones in the United States. Note how the cyclone tracks converge toward the northeastern part of the country. Cold anticyclones have a dual origin. Those which move south from northern Canada (*cP* air) bring severe cold in winter; those from the Pacific Northwest (mo[illegible] *mP* air) bring only moderate cold. Many of the l[illegible]r are in the nature of weak highs separating traveling cyclones. Warm anticyclones are not shown.

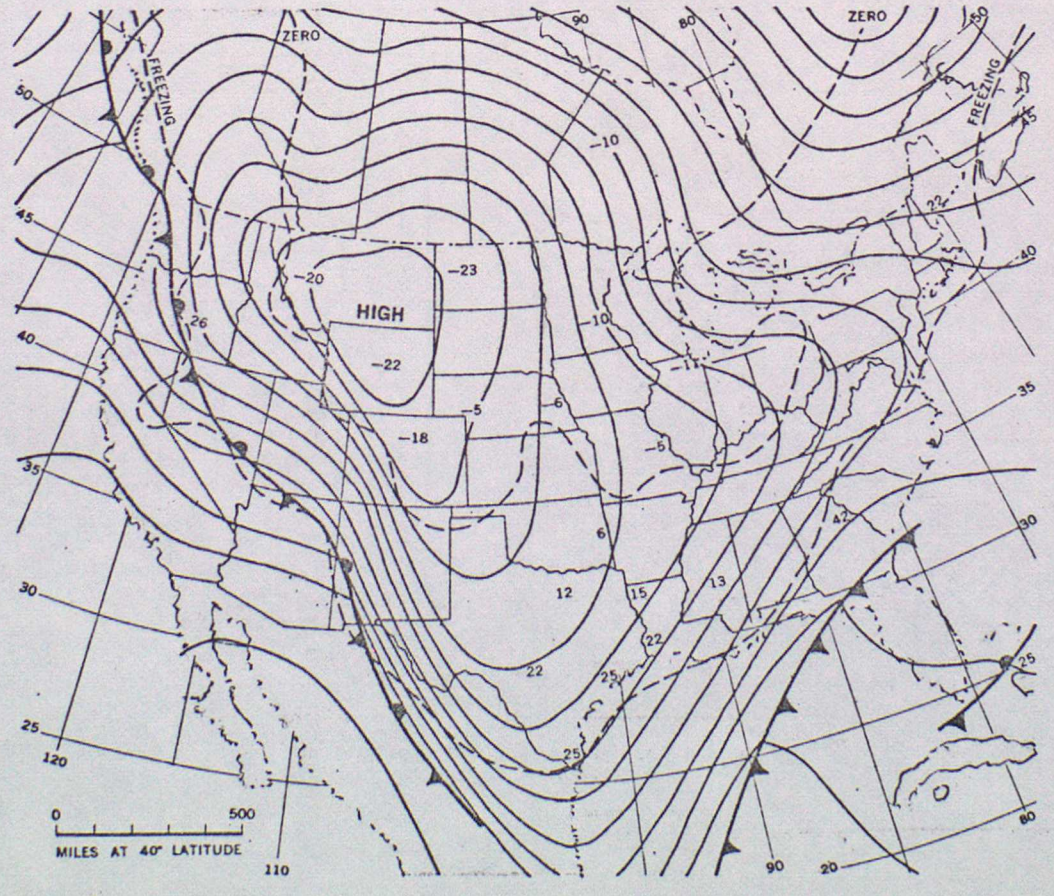


Figure 2.19 Polar-outbreak cold anticyclone advancing rapidly southward as a mass of unusually cold and stable *cP* air. On this occasion Minot, North Dakota, had a minimum temperature of 23° below zero, Denver −18°, and Jackson, Mississippi, 15°. Below-freezing night temperatures were common along the margins of the Gulf of Mexico. (*From Weatherwise, June, 1962.*)

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LARGE-SCALE INTERANNUAL VARIABILITY OF CLIMATE

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ABSTRACT

Climate is discussed in terms of research into its interannual variability on large spatial scales. Some of the statistical properties of this variability are presented. Several examples of hemispheric-scale fluctuations are used to elucidate the physical mechanisms at work.

1. Introduction

Variability is an essential feature of climate. To regard the normal alone as an adequate expression of climate would be a serious error; it is better to regard current climate as the entire ensemble of recently observed atmospheric conditions. The possibility that the climate itself changes on time-scales of decades or longer will be discussed in other lectures: the present discourse is mainly concerned with atmospheric variations on time-scales of 1 to 5 years. Such variations may be regarded as expressions of the existing climate, rather than as changes of climate: there is, of course, no real dividing line between inter-annual fluctuations and climatic change.

Most of the research into large-scale interannual variability in the atmosphere has, until recently, been of an empirical nature, not only because of the need to establish the facts before setting up hypotheses and models, but also because of the difficulty in formulating relatively simple working hypotheses which not only fit the facts but also represent realistically the complex physics and dynamics of the atmosphere and ocean. A further limitation has been the impossibility of integrating complex numerical models through simulations of many years. This account will begin empirically by illustrating the magnitude of interannual variability, the statistical distributions to which it appears to conform, and the degree of regularity of the variability in time and space; but then an attempt will be made to examine the underlying causes of interannual variations, in the context of the tropical stratospheric quasi-biennial oscillation and of the large-scale tropical fluctuations known as the Southern Oscillation.

2. Illustrations of interannual variability

Variability of climate in terms of temperature and rainfall is a natural consequence of variability of the atmospheric circulation, which will therefore be the main subject of this section.

a. The magnitude of climatic variability

The magnitude of the variability inherent in the atmospheric circulation is illustrated by Figures 1a and 1b which show the normal January mean sea level pressure pattern and the standard deviation of individual January averages. It is clear, from a comparison of the normal pressure gradients with the standard deviations, that the variability is sufficient to occasionally reverse major features, such as the midlatitude westerlies in the Atlantic, for whole months. The maxima of variability are near the subpolar lows and have been found to be related not only to statistical scatter in the depths of individual depressions, but also to variability on time-scales of weeks, associated with the occurrence of major interruptions of the midlatitude westerlies, i.e. blocking. The magnitude of the interannual variability of circulation is exemplified by the fact that the Northern Hemisphere atmospheric annual mean angular momentum varies by up to 25% (Rosen et al 1976).

Circulation variability entails variability in cloud cover, resulting in variability in the Earth's long-wave radiation budget as documented by Heddinghaus and Krueger (1981). The effects of circulation variability on local temperature and rainfall are a complex function of the Earth's geography and will not be discussed here.

An important component of climatic variability is in the sea-surface temperature. Figures 2a and 2b show the February normal of sea-surface temperature and the standard deviation of individual February averages. In this case, the variability is generally insufficient to reverse the North-South temperature gradient. However, the variability of sea-surface temperature in some low-latitude regions such as the tropical East Atlantic (Rowntree 1976), and the equatorial East Pacific (Bjerknes 1966, 1969), appears to be physically linked to the atmospheric variability, both locally and far afield. The equatorial East Pacific is discussed below in the context of the Southern Oscillation. The importance of the midlatitude North Pacific has also been stressed (e.g. Namias 1978).

Attempts to simulate atmospheric variability with 3-dimensional dynamical models have hitherto been thwarted by the computational difficulty and expense of including an ocean which interacts realistically with the atmosphere. However, in contrast to the above mentioned results on sea-surface temperature, Manabe and Hahn (1981) found that omission of interannual sea-surface temperature variability from their general circulation model resulted in an underestimation of atmospheric variability only in the tropics, where the oceanic aspects of the Southern Oscillation could not be represented.

b. The statistical distribution of climate

The possibility that uninterrupted westerly flow, and blocked flow, mentioned above, are two alternative equilibrium states for the midlatitude atmosphere (cf Charney and DeVore (1979)), is suggested by the discovery that some measures of the circulation show skewness or even bimodality in their frequency distribution. The example in Figure 3, Azores minus Iceland mean sea level pressure for winter months, is skew and slightly bimodal. The bimodality was found to be greater in periods when blocking was more frequent, e.g. 1940-81. The skewness is reflected in the statistical distribution of winter temperatures in Central England, which are strongly controlled by the Atlantic flow. The Azores minus Iceland mean sea level pressure distribution for the other seasons is more nearly Gaussian.

The monthly averaging process applied to daily winter European temperatures also fails to produce a Gaussian frequency distribution of these temperatures because of the extreme coldness of easterly flows. By way of contrast, Parthasarathy and Mooley (1978) found that Indian summer monsoon rainfall conformed to a Gaussian distribution. Individual daily rainfall values are unlikely to approximate to a Gaussian distribution because a large frequency of dry days (zero rainfall) is likely to be superimposed on a J-shaped distribution for days with rain: but the daily Indian values must have satisfied the conditions in which the central limit theorem applies, so that averaging produced a more Gaussian result.

c. Quasi-periodicity

The best-known example of an almost periodic interannual variation is the quasi-biennial oscillation of wind direction and speed and temperature in the lower stratosphere in the tropics. The monthly mean 30mb winds over Canton Island (Pacific) up to 1967 then over Gan (Indian Ocean) are shown in Figure 4.

It is possible to form such a composited series from two widely-separated stations only because the quasi-biennial oscillation occurs on a global scale in the tropics with very little time-lag between different longitudes. The period of the oscillation is not constant: it varies from less than 2 years to about 3 years, and averages 2 years and 3 months. Periods near either 2 or 3 years tend to be favoured, and there are favoured seasons for changeover of the wind direction (Parker 1976). There is also downwards phase propagation, with, for example, changes in the 50mb wind vector lagging by several months behind changes in the 30mb wind vector.

The discovery of the tropical statospheric quasi-biennial oscillation has been followed by very widespread reports of quasi-biennial behaviour in local tropospheric climate and circulation parameters. Parthasarathy and Mooley (1978) found a quasi-biennial oscillation in Indian Monsoon rainfall, but this was not strong and regular enough to be used for annual forecasting. Ebdon (1975) documented differences in the mean Northern Hemisphere midlatitude surface pressure and 500mb geopotential patterns, and in zonal winds aloft, according to the phase of the tropical quasi-biennial oscillation. Figure 5 illustrates the mean sea level pressure anomalies for Julys with a westerly tropical quasi-biennial wind mode. The anticyclonic anomaly near the British Isles was found to be statistically significant at about the 5% level. In addition, not only Ebdon (1975) but also Angell and Korshover (1977) observed that the circumpolar upper-tropospheric vortex was smaller during the westerly phase of the tropical quasi-biennial oscillation, particularly in winter (Figure 6), with an associated increase of tropospheric temperature of several tenths ^{o}C in midlatitudes. Consequent effects on, for example, summer temperature and sunshine in the United Kingdom, were noted by Ebdon. Furthermore, Holton and Tan (1980) have demonstrated that there is a 50mb and 1000mb quasi-biennial oscillation in the geopotential and temperature of the northern polar regions. The two levels are in phase but the effect is weaker at 1000mb. It must be emphasised that these observations of quasi-biennial oscillations outside the tropical stratosphere have only been associated statistically with the tropical stratospheric quasi-biennial oscillation at this stage. Physical associations have not been established.

d. Teleconnections

Interannual variations in the atmosphere are sometimes coherent on very large scales. In fact, events at locations in differing regions of the globe have often been found to be statistically connected in the sense that opposite effects tend to occur at distant places (the effects are negatively correlated). An important example of such a "teleconnection" is the Southern Oscillation, documented by Walker and Bliss (1932) as an inverse but non-periodic pressure variation on interannual time-scales between the East Pacific Ocean and the East Indian Ocean at low latitudes. The major Northern Hemisphere long-distance relationships are summarized for winter in Figure 7, from Wallace and Gutzler (1981), in which negative correlations are indicated between locations at opposite ends of the arrows. The variations are not periodic. Of particular interest are the North Atlantic Oscillation, which is an inverse variation of the pressure in the Azores anticyclone and the Iceland low, a similar effect in the western North Pacific, and a negative correlation between pressure in the Aleutian low and over the northern Rockies, forming a link in a chain of events tenuously connecting the tropical Pacific with the Atlantic. The fluctuations in the Iceland low result in inverse variations ("seesaw") in winter temperature in Greenland and Northern Europe (Van Loon and Rogers (1978)). The pattern of teleconnections over the East Pacific and America in Figure 7b is reminiscent of the wavetrains resulting from subtropical thermal and orographic forcing in the 5-layer baroclinic model of Hoskins and Karoly (1981), whose results may also throw light on the relationships between the Southern Oscillation and interannual variability in midlatitudes, discussed in the next section.

3. Hypotheses for interannual variability

There are at least three classes of explanation of atmospheric interannual variability: interactions within the atmosphere, whereby short-period fluctuations create and maintain longer-term equilibria; thermal interactions with the oceans and cryosphere (ice and snow); and external forcings. The latter group of theories will not be discussed here: it includes the influence of variations in solar constant, which has however hitherto been very difficult to measure to the required accuracy; and the influence of volcanic dust, which will be considered in the lecture on climatic fluctuations. In the present lecture, internal atmospheric interactions will be discussed in the context of the tropical stratospheric quasi-biennial oscillation; and atmosphere-ocean interactions will be illustrated by an examination of the Southern Oscillation.

a. Interactions within the atmosphere: the quasi-biennial oscillation

Holton and Lindzen (1972) have formed a hypothesis that the downward-propagating quasi-biennial oscillation in the tropical lower stratosphere is driven by vertical convergence of the momentum flux of radiatively dissipating, vertically propagating, short-period (5-15 days) waves excited in the upper troposphere. The radiative dissipation results from reduction of the amplitude of the temperature cycle of the wave by Newtonian cooling.

Imagine an atmosphere initially at rest until waves of a type which propagate westerly momentum upwards are excited near the tropopause, as a result of e.g. geographically irregular tropospheric heating. Equatorial Kelvin waves have the appropriate properties, propagating eastwards as well as upwards. The waves will propagate upwards with weak attenuation up to near 30 km (10mb) where, because the radiative dissipation increases with height, their amplitude will decrease rapidly with height, leading to a rapid reduction with height in the upward westerly momentum flux associated with the waves, and hence to westerly acceleration of the mean flow. As a result there will be a reduction of the eastward phase speed relative to the mean flow. If the acceleration of the mean flow is slow compared with the accelerations involved in the wave motions, linear approximations will apply and it can be assumed that the vertical phase speed of the waves will be reduced in proportion to the horizontal phase speed. The reduced vertical phase speed will allow more time for Newtonian cooling to reduce the wave amplitude, in other words there will be increased radiative dissipation of the waves where they encounter the westerly flow, i.e. in the shear zone beneath the strongest westerly flow. The resulting decrease of wave amplitude with height will lead to convergence of the wave momentum flux at these lower levels. The momentum will no longer be available for the upper levels, where the acceleration will therefore decrease. However the increased acceleration at the lower levels will lower the shear zone, further lowering the level at which most of the momentum flux converges. Eventually a westerly shear zone will descend to near the tropopause where a balance will be established between convergence of wave momentum and vertical diffusion of momentum into the troposphere.

Now imagine that waves which propagate easterly momentum upwards are also excited near the tropopause; such waves are also observed in the tropics. These waves will be able to propagate through the westerly shear zone but, as a result of radiative dissipation at higher levels, their easterly momentum flux will converge there, so that, in a manner similar to that described above, a downward-propagating easterly shear zone will follow the westerly shear zone. Eddy diffusion will permit the easterly shear zone to propagate to the tropopause without leaving a shallow layer of westerlies.

Holton and Lindzen illustrated their hypothesis with a one-dimensional analytical model, which also included an imposed climatological mid-stratospheric tropical semiannual oscillation of zonal wind. Their results are presented in Figure 8. The donward propagation and periodicity (27 months) are similar to what on average is observed, but the periodicity could have been varied by altering the radiative dissipation rates, or by changing the wave amplitudes at the lower boundary. However, their values for these parameters were all compatible with observations, lending credence to their theory. The semiannual oscillation may be responsible for the synchronous nature of the quasi-biennial oscillation throughout the tropics, because, by affecting wave phase speeds and thus momentum convergence in the 30-40 km layer, it could limit the possible times of onset of new shear zones. However, the semiannual oscillation is not essential to Holton and Lindzen's theory and they did not regard it as determining the time-scale of the quasi-biennial oscillation.

Plumb (1977) has suggested modifications to Holton and Lindzen's theory. In particular, because the wave amplitude and therefore the momentum convergence at a given level are controlled by the mean flow at lower levels, he found the vertical variation of radiative cooling to be unnecessary. Also Plumb found that the destruction of the lowest zone was a result of eddy diffusion of momentum from the zone of reverse winds immediately above, rather than eddy diffusion into the troposphere.

The mechanisms by which the tropical lower stratospheric quasi-biennial oscillation effects the rest of the atmosphere are still uncertain. It is possible that the associated temperature changes near the tropical tropopause may affect the strength of convection, and thereby influence the Hadley cell and subsequently the rest of the circulation (Folland, private communication).

b. Atmosphere-ocean interactions: the Southern Oscillation

The Southern Oscillation has already been referred to as an inverse variation of pressure between the Indian and Pacific Oceans in the tropics. Figure 9 shows the global scale of the phenomenon, and Figure 10c, in which a positive index denotes above-normal pressure over the tropical southeast Pacific, shows that the fluctuations are irregular in time but occur on time-scales of the order of several years. However, Figures 10a, b and d also show the involvement of winds, sea-surface temperatures and rainfall in the equatorial Pacific. When pressure is higher than normal over the tropical southeast Pacific, there is enhanced easterly flow along the equator in mid-Pacific, accompanied by reduced sea-surface temperatures and reduced rainfall. There is also an enhanced westerly return flow aloft (Julian and Chervin 1978), completing a direct circulation in an equatorial plane, known as a "Walker Cell". According to Bjerknes (1966), the reduced sea-surface temperatures are a result, not mainly of westward advection of water from the cold Humboldt Current, but largely of enhanced upwelling of water as a result of the wind-induced Ekman drift: persistent easterlies result in poleward motion of surface water in geostrophic equilibrium, and the consequent equatorial divergence can only be balanced by equatorial upwelling. The rainfall in the equatorial mid-Pacific is reduced, partly because of suppression of local convection over the cooler ocean, and partly because of subsidence associated with the large-scale vertical circulation (Bjerknes 1969). At the opposite stage of the Southern Oscillation cycle, pressure is lower than normal over the tropical southeast Pacific, with reduced mid-Pacific equatorial easterlies, increased mid-Pacific rainfall, and increased eastern and mid-Pacific sea-surface temperatures. Horel and Wallace (1981) have also shown that the entire tropical troposphere is warmer at this stage of the Southern Oscillation. It is thus an important phenomenon on a global scale.

Bjerknes (1969) considered either stage of the Southern Oscillation to be self-amplifying. For example, an increase in the pressure difference is associated with an increase in the easterlies, leading to a reduction of the sea-surface temperatures in the central and eastern equatorial Pacific, and hence to an increase of east-west temperature contrast; so that the Walker direct cell is strengthened, giving a further increase in pressure difference, and stronger winds, and so on. But in his 1969 paper Bjerknes was unable to specify the means of turnabout between opposite stages of the Southern Oscillation. In his 1966 paper, however, he had proposed the well-known hypothesis that an anomalously warm ocean surface would provide extra energy to the north-south (Hadley) circulation at that longitude. It would follow (Julian and Chervin 1978) that only the subtropical westerly jet but also the low-latitude low-level easterlies would increase, the equatorial ocean would cool, and so the Southern Oscillation would change to its opposite phase. However, the cool ocean would now be a reduced energy source for the Hadley circulation, which would weaken, giving reduced low-latitude easterlies, and a warming of the equatorial ocean, so that the Southern Oscillation would have come full circle.

If the above chain of reasoning were correct, there would be a lag, on the time-scale of atmospheric circulation changes, i.e. days or weeks, between the time of warmest ocean and the time of strongest easterlies. Figure 10 does not support this: the strongest equatorial easterlies appear to occur towards the end of periods with cold ocean. Wyrtki (1975) therefore proposed a different hypothesis to explain the changeover of the Southern Oscillation from positive index (strong winds, cold ocean) to negative (warm ocean). On the basis of sea - level observations from coastal and island stations, he suggested that the increased easterlies in the cold-ocean phase can strengthen the South Equatorial Current and thereby maintain an anomalous slope of sea-level, upwards towards the west. But within a few months of a weakening (for an unspecified reason) of the winds, the accumulated warm water will move east from the west Pacific, and the thermocline will be lowered, so that the remaining winds will be unable to cause upwelling of cold water, even though there is some evidence that upwelling is at least as intense at this stage (Anderson, 1981). Thus the warm-ocean phase will be established. The warming only occurs during Northern winter, and Wyrtki also showed that the same phenomenon occurs in a less intense form during every Northern winter. He did not attempt to explain the mechanism of reversion to the cold-ocean phase of the Southern Oscillation. Neither does Wyrtki's hypothesis appear to explain the fact (Horel and Wallace 1981) that the equatorial central Pacific warms several months later than the Peruvian coastal waters. Barnett's (1977) statistical study lends support to Wyrtki's hypothesis so far as the eastern equatorial Pacific is concerned, and then favours westward advection of less cool water, and reduced wind-induced equatorial upwelling, as reasons for the later mid-Pacific warmth.

Despire the inadequacy of Bjerknes' hypothesis in explaining the turnabouts of the Southern Oscillation, it has been shown, in agreement with his hypothesis, that the annual mean strength of the Northern Hemisphere subtropical jet stream in the longitude range of interest increases as the eastern equatorial Pacific warms (Julian and Chervin 1978). In addition (Van Loon and Rogers 1981) the characteristics of the winter Northern Hemisphere midlatitude circulation are changed, with increased heat transport in the quasistationary eddies, lower mean temperature between 30°N and 60°N, and heat transport by transient eddies reduced north of 45°N but increased south of 45°N. Similar adjustments of the midlatitude long waves appear in the steady-state linearised primitive equation model of Hoskins and Karoly (1981) when a subtropical forcing (heat source) is introduced. These and other theoretical results also indicate that strong teleconnections to midlatitudes are possible only when the upper tropospheric westerlies extend over the low-latitude heat source. For the Northern Hemisphere mid-Pacific this condition is only fulfilled in the winter half-year, and the observational evidence indeed suggests that the strong teleconnections with midlatitudes occur only in winter.

4. Conclusion

Future investigation of large-scale interannual variability of climate requires observational and modelling studies of atmospheric and oceanic aspects. Modelling studies may in some cases be in simple forms for testing specific hypotheses, as in the quoted work on the quasi-biennial oscillation: but many phenomena appear to require fully-fledged models for adequate representation. In the particular case of the Southern Oscillation, an interactive ocean-atmosphere model would be of considerable value in investigating what has turned out to be a highly complex and intriguing phenomenon.

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Figure 1a

NORMAL OF MONTHLY MEAN SURFACE PRESSURE (DATA FOR 1900-1980)

JANUARY (mb)

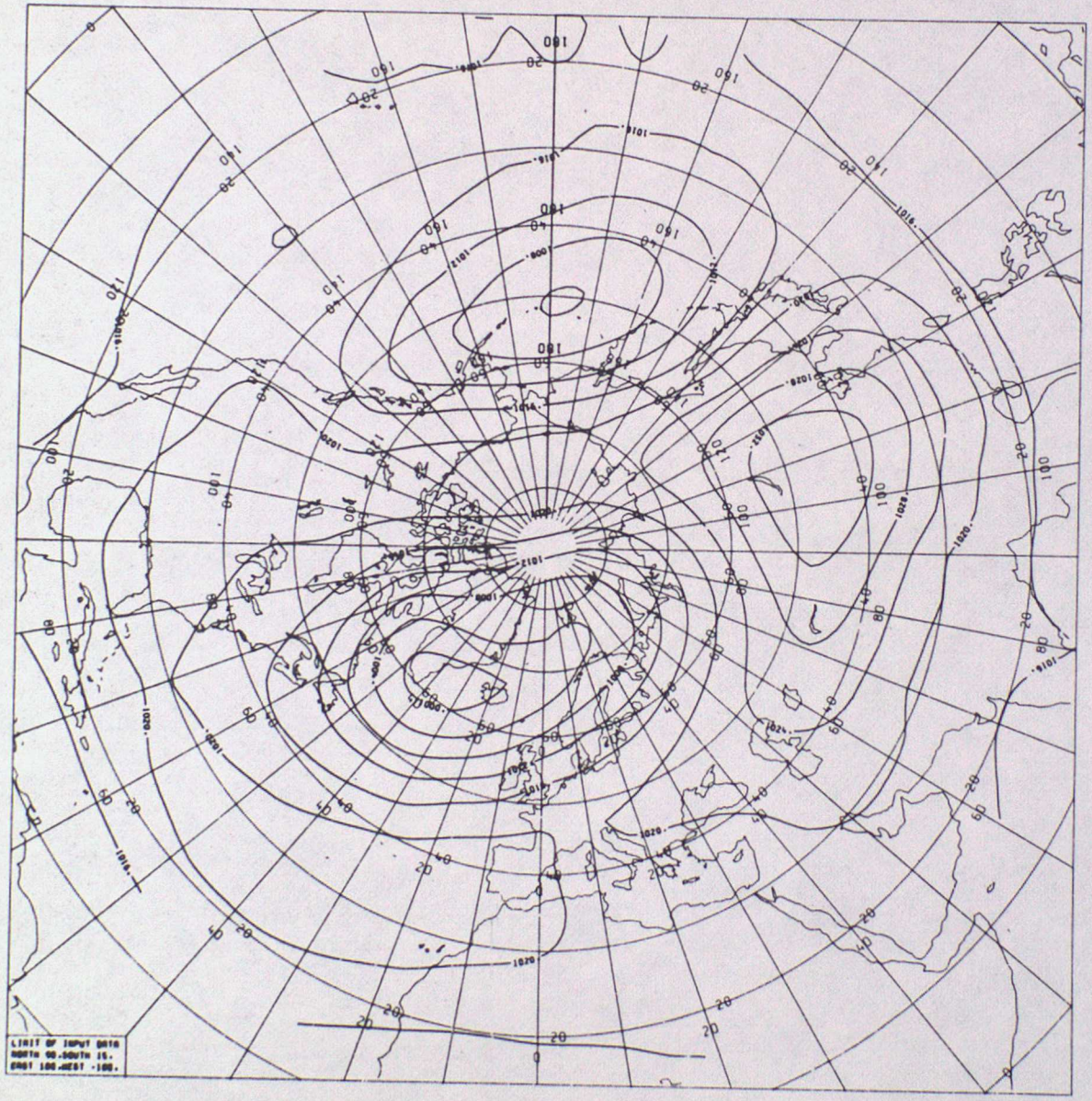


Figure 1b

STANDARD DEVIATION OF MONTHLY MEAN SURFACE PRESSURE (DATA FOR 1900-1980)

JANUARY (mb)

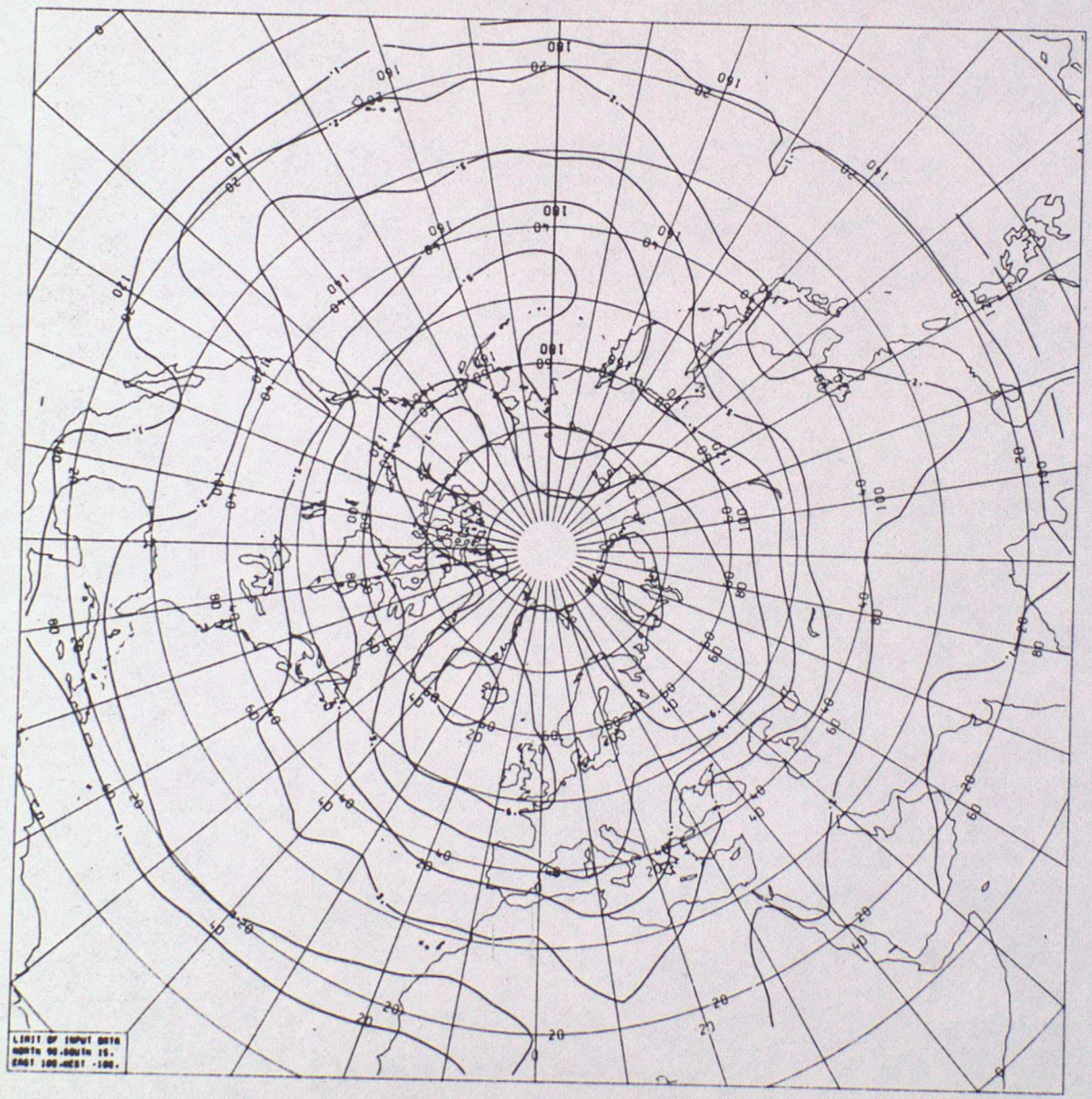


Figure 2a

NORMALS OF SEA SURFACE TEMPERATURES (°C)
FEBRUARY DATA: 1900 - 1960 ATLANTIC AREA
1931 - 1960 OTHER AREAS

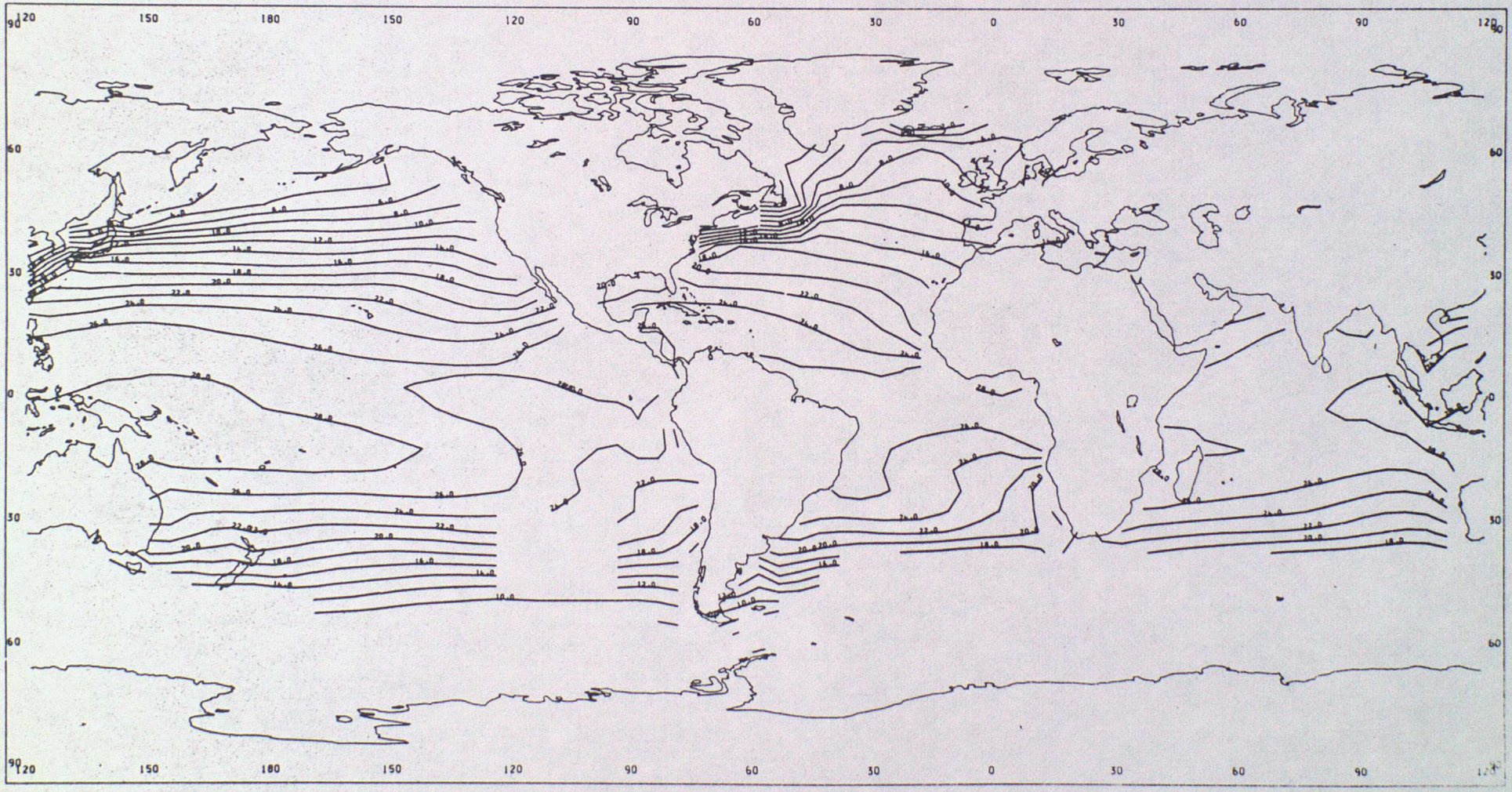
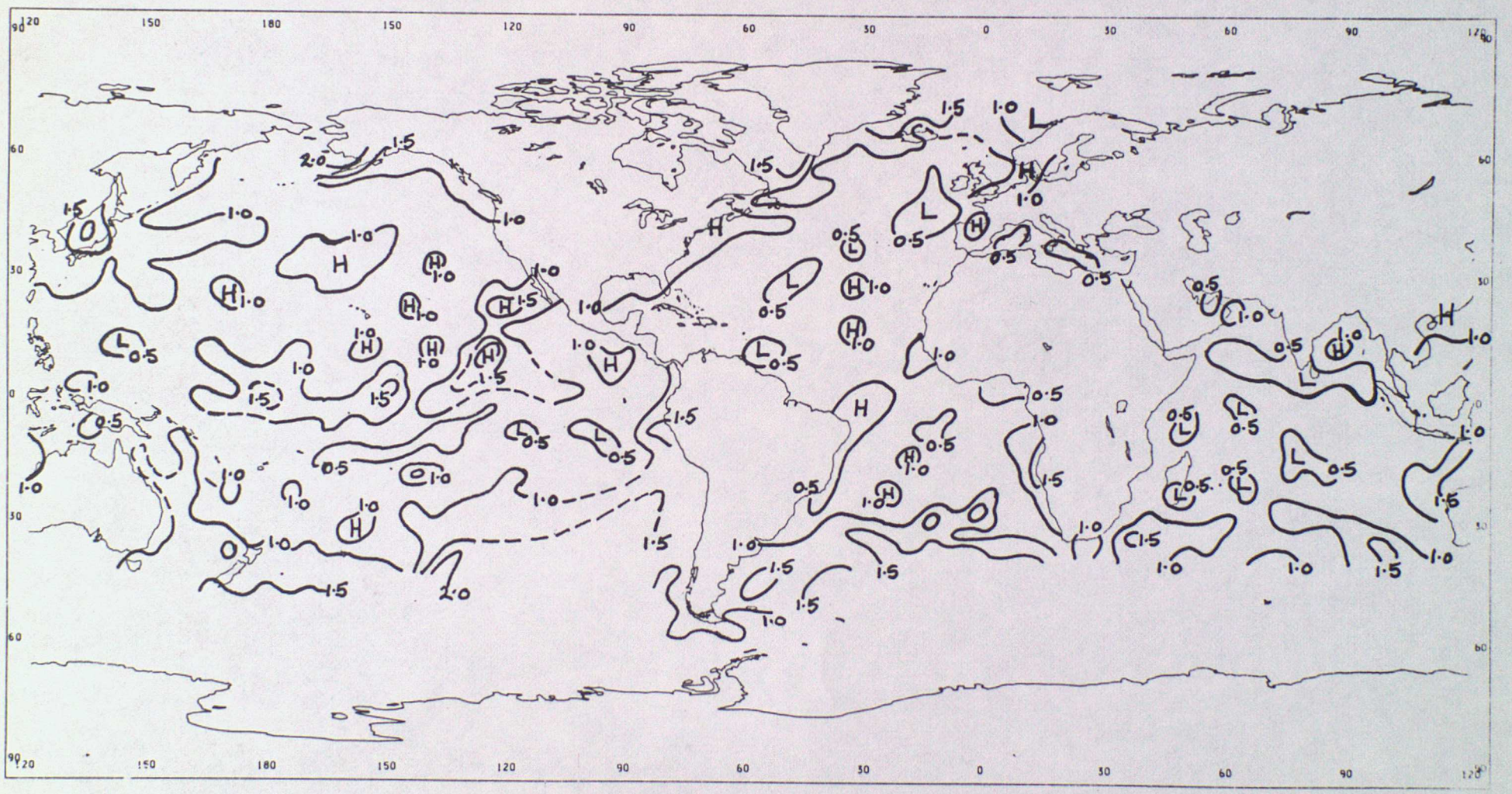


Figure 2b

STANDARD DEVIATION OF SEA SURFACE TEMPERATURE DATA
ABOUT MONTHLY NORMALS (°C)
FEBRUARY DATA: 1900 - 1960 ATLANTIC AREA
1931 - 1960 OTHER AREAS



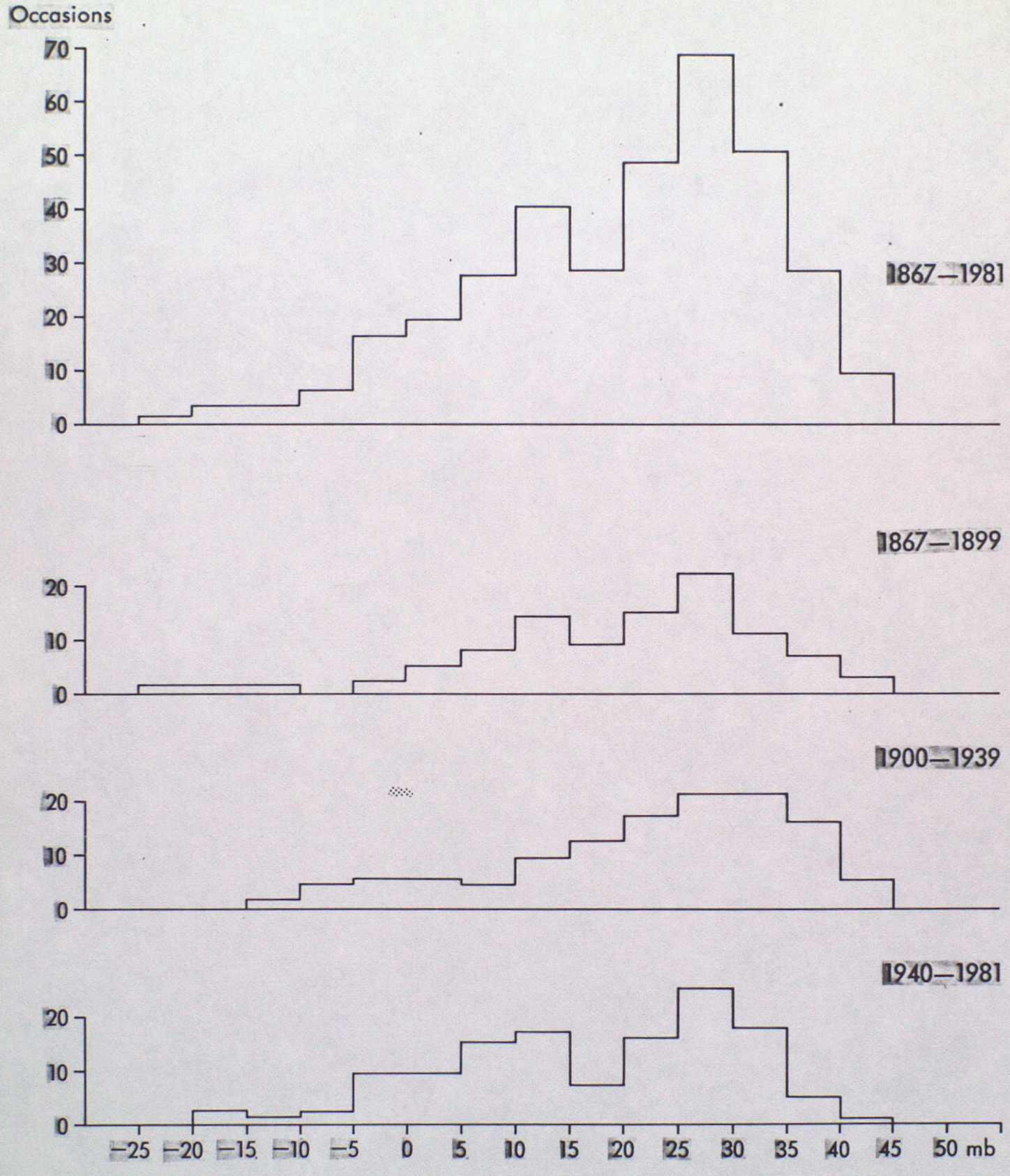


Figure 3. Ponta Delgada (Azores) minus Stykkisholmur (Iceland) mean sea level pressure, winter months (up to Dec. 1981).

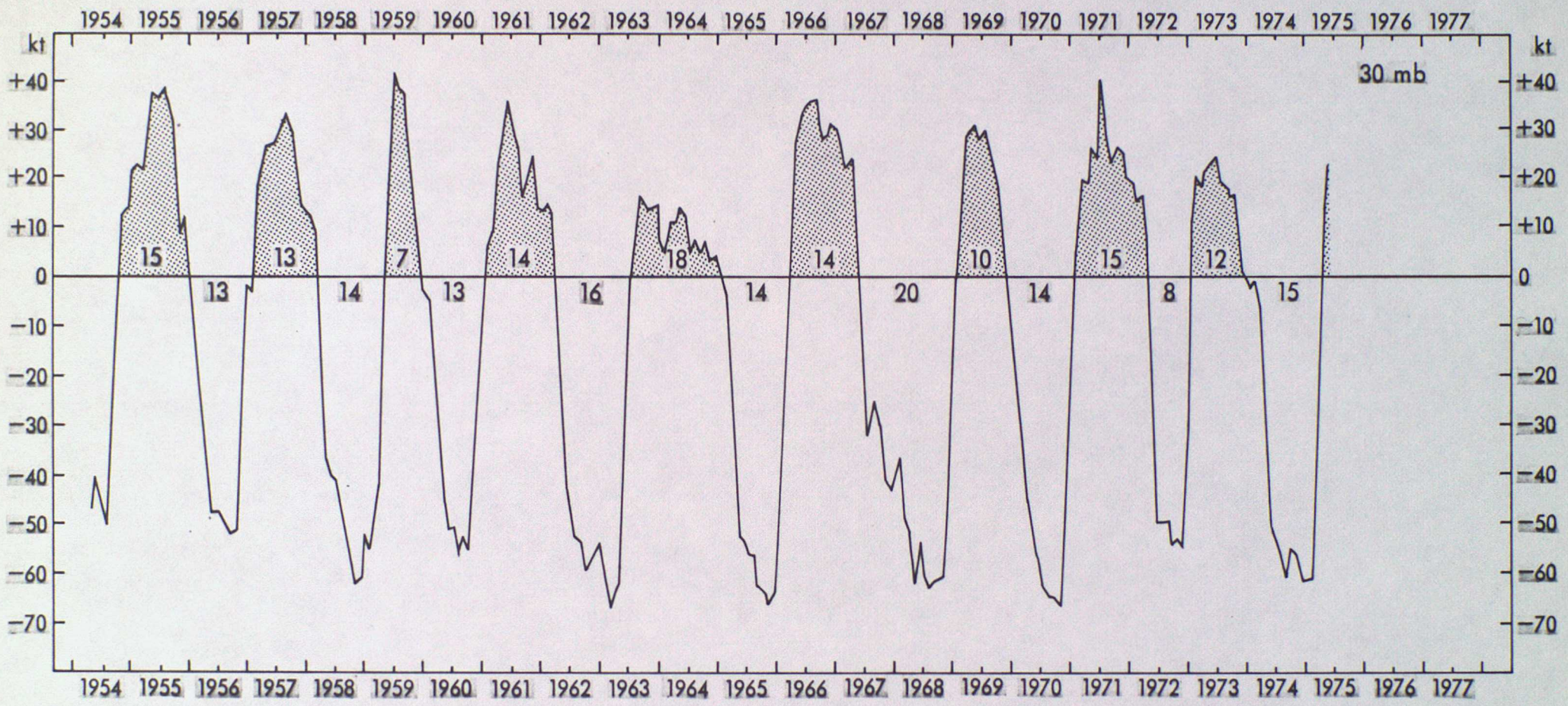


Figure 4. 30-mb monthly mean zonal wind components at Canton Island/Gan. After Ebdon (1975). Components towards the east are positive and stippled. Figures along the zero line indicate the duration, in months, of westerlies or of easterlies.

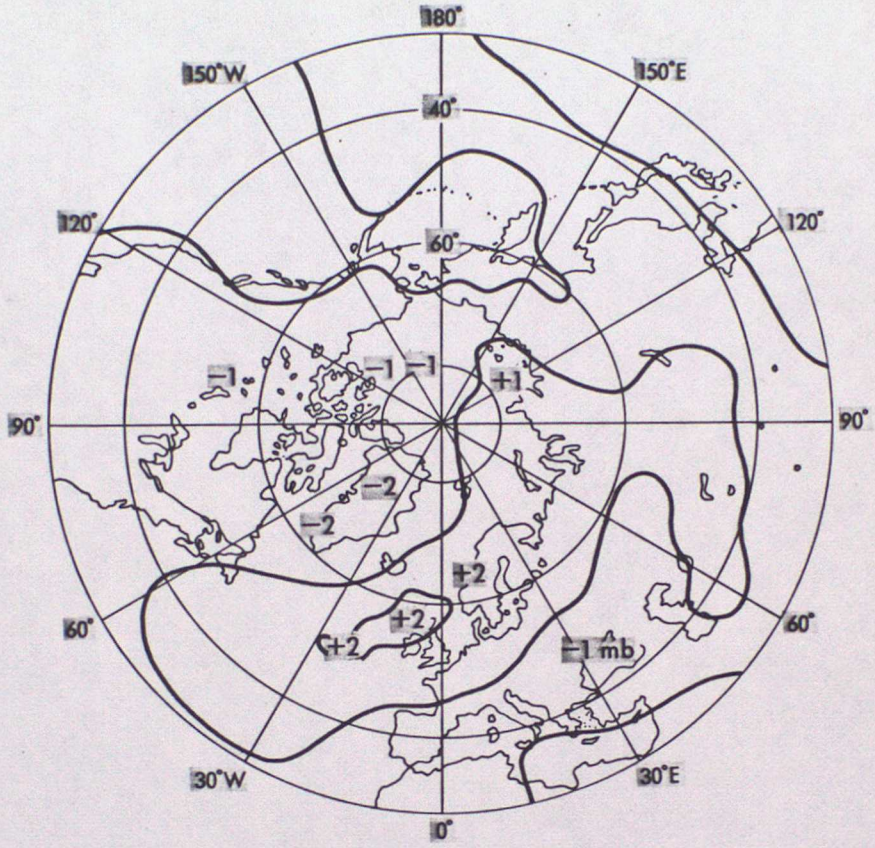


Figure 5. Surface pressure anomaly from 1951—70 average, July.
After Ebdon (1975).
Mean of eight Julys with westerly phase of QBO at 30 mb.

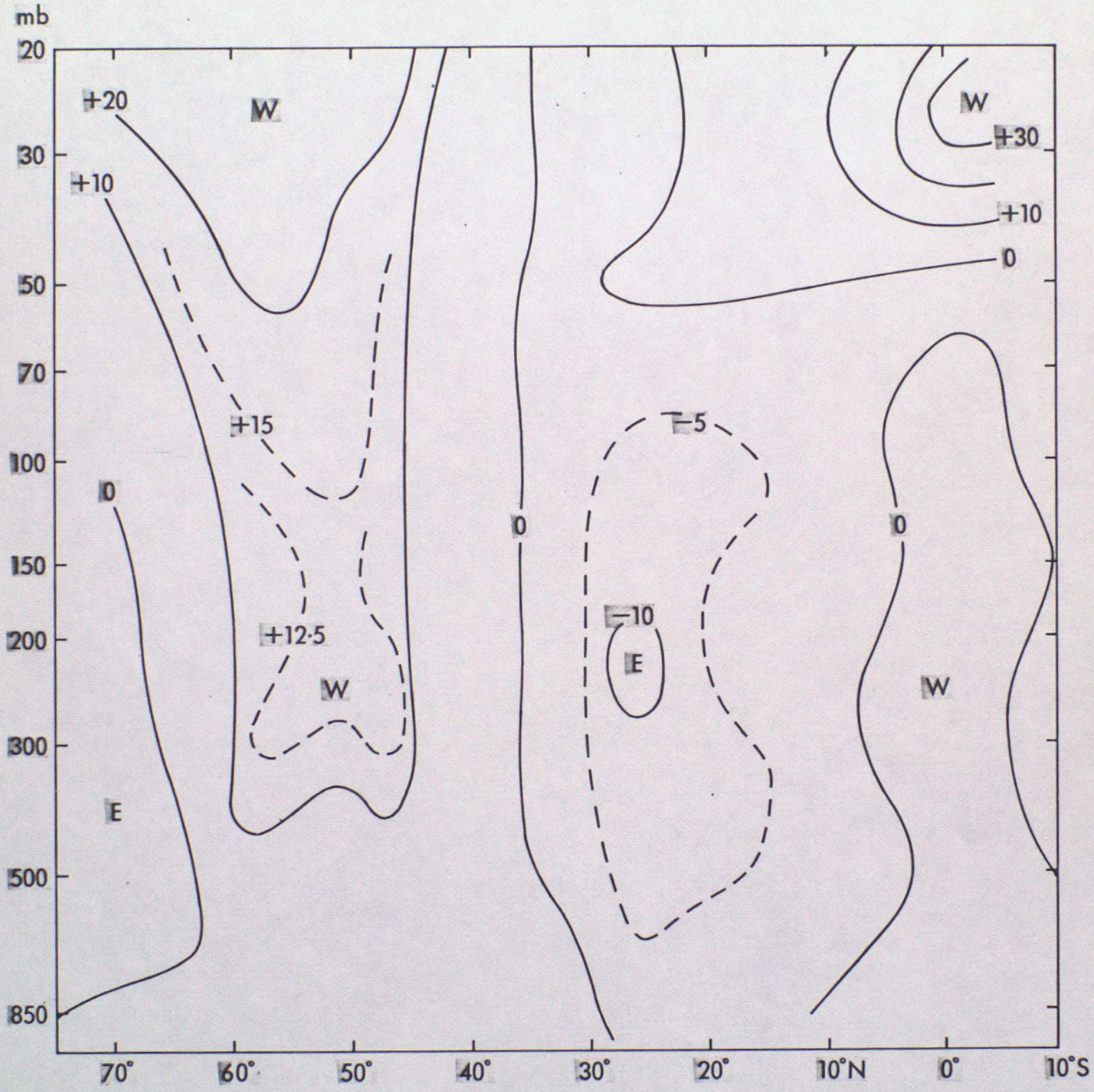


Figure 6. Vertical cross-section at 80°W showing differences between zonal component during westerly and easterly phases of the QBO at 30 mb in January. After Ebdon (1975). Speeds are in metres/second.

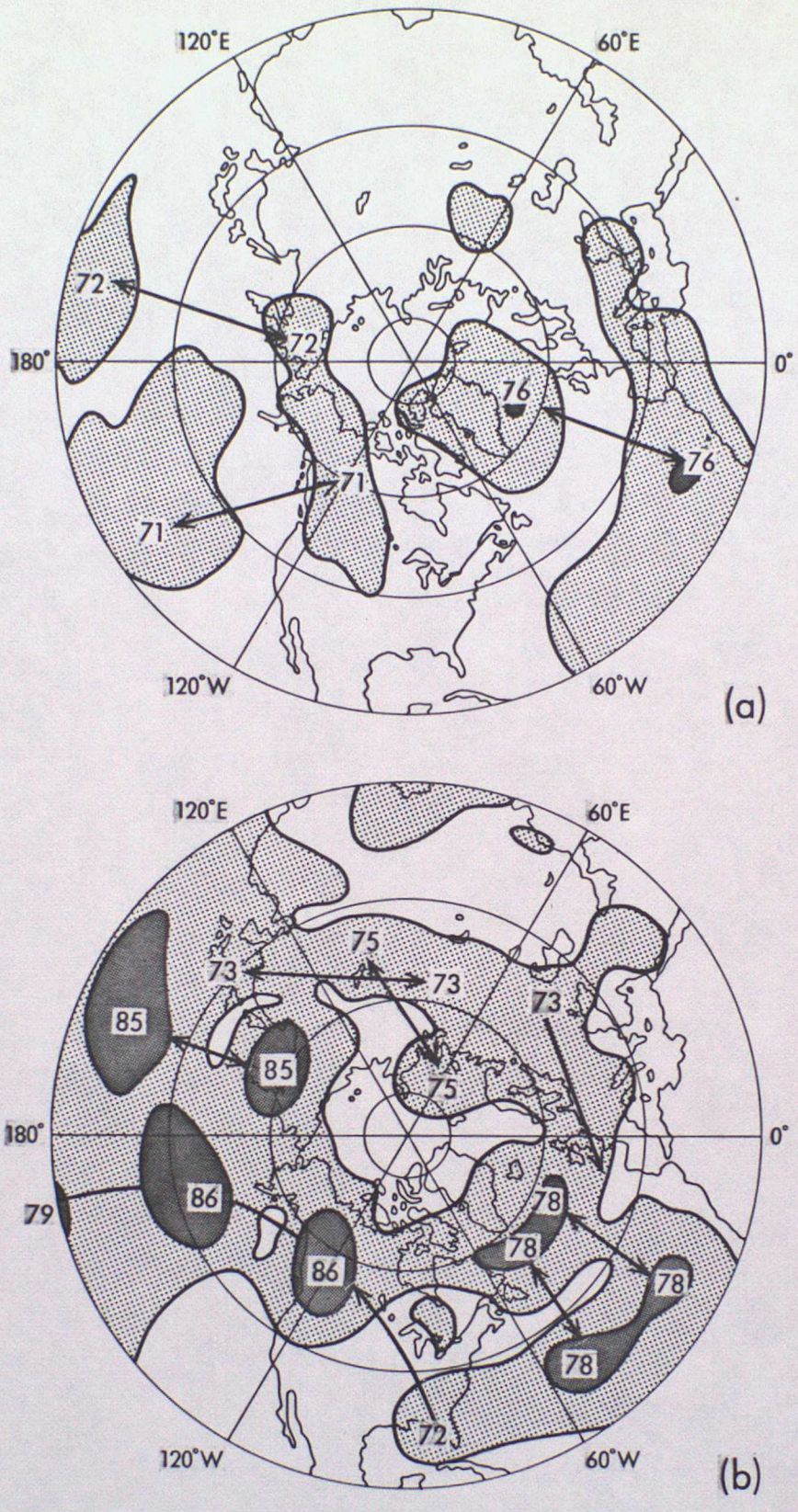


Figure 7. Strongest negative correlation ρ_i with any point, plotted at the base grid point for (a) sea-level pressure and (b) 500 mb height. Based on 45 winter months (Decembers, Januarys, and Februarys for winters 1962—63 through 1976—77). Negative signs have been omitted and correlation coefficients multiplied by 100. Regions where $\rho_i < 60$ are unshaded; $60 \leqslant \rho_i < 75$ stippled lightly; $75 \leqslant \rho_i$ stippled heavily. Arrows connect centres of strongest teleconnection. (After Wallace and Gutzler 1981).

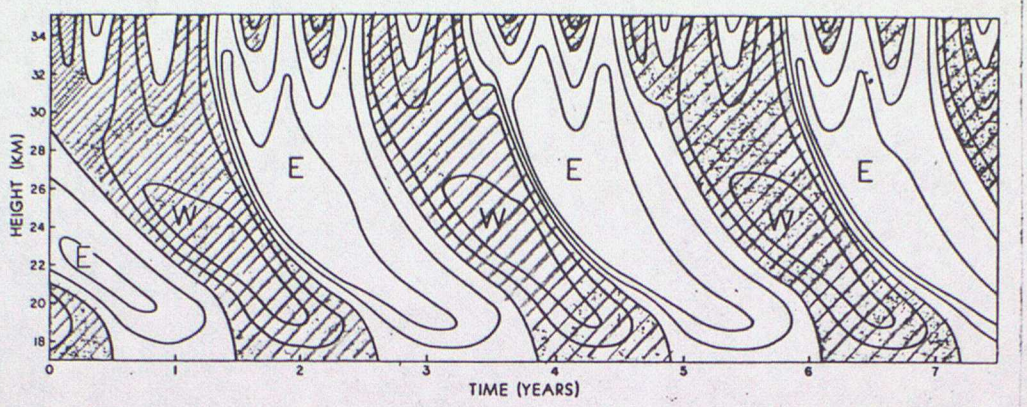


FIG. 8. Time-height cross section of mean zonal wind. Contours have been drawn at 10 m sec^{-1} intervals. Regions of westerly flow have been shaded.

From the model of Holton and Lindzen (1972).

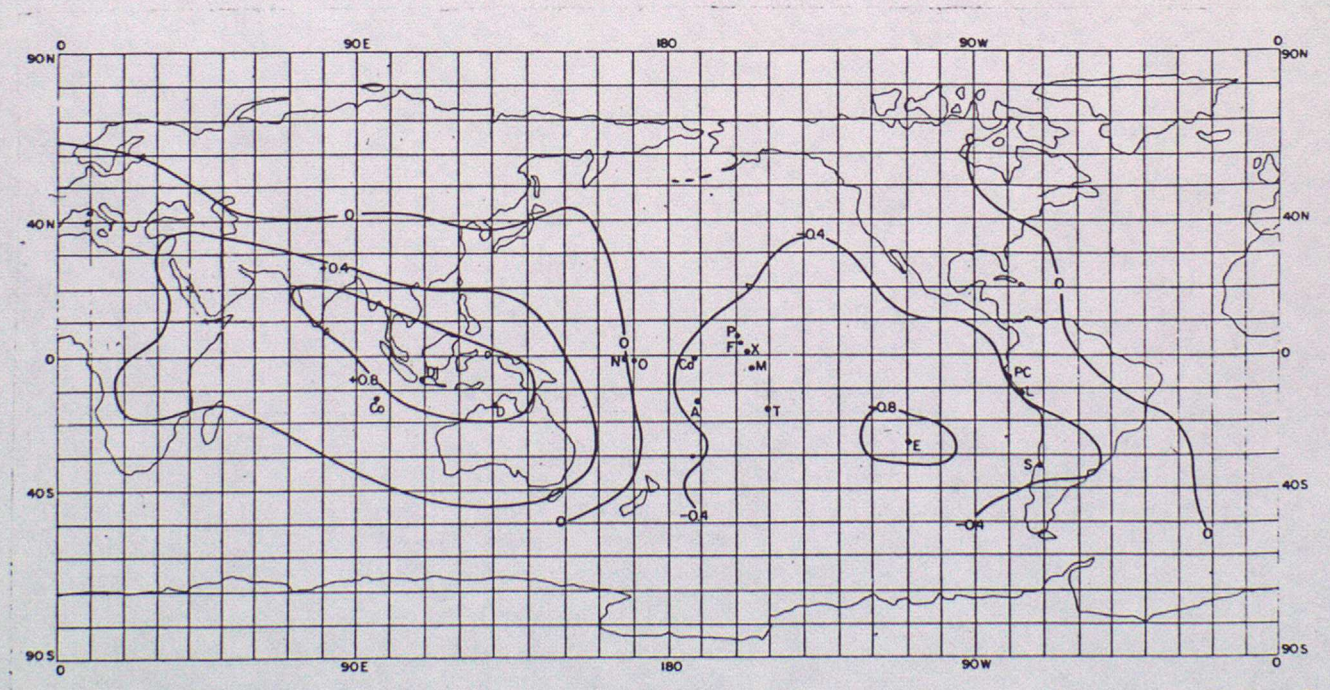


FIG. 9. Schematic map (after Berlage, 1966) showing isopleths of correlation of monthly mean station pressure with that of Djakarta, Indonesia (Dj). Other localities shown are Cocos Island (CO), Port Darwin (D), Nauru (N), Ocean Island (O), Palmyra (P), Christmas Island (X), Fanning (F), Malden Island (M), Apia, Samoa (A), Tahiti (T), Easter Island (E), Puerto Chicama (PC), Lima (L) and Santiago (S).

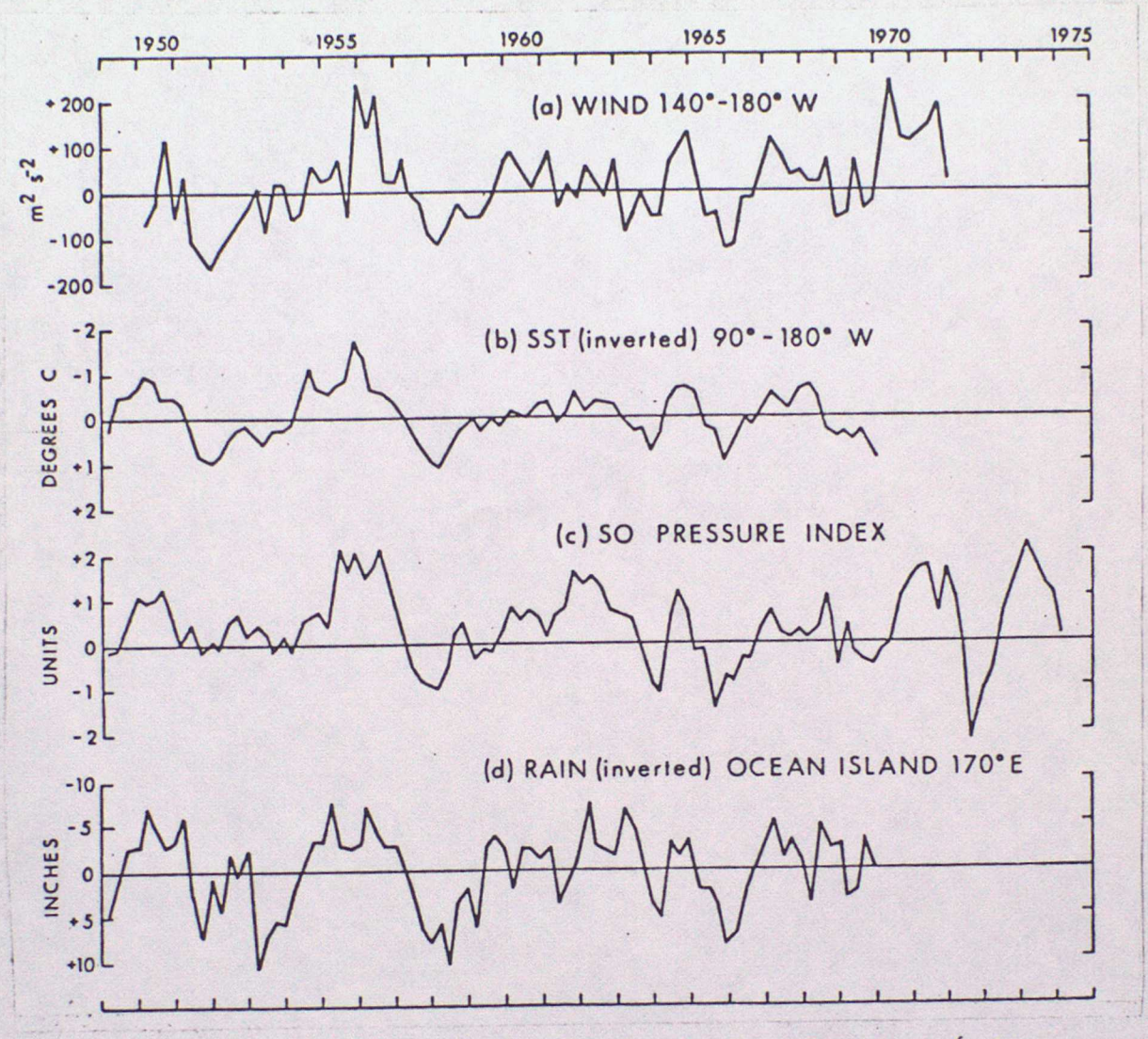


Figure 10. Parameters of the Southern Oscillation (after Wright 1977).

a) Zonal component of wind stress, 4N-4S 140-180W, seasonal anomalies relative to 1950-1969 mean (after Wyrtki 1975);

b) Sea-surface temperatures 10N-10S, 90-180W, seasonal anomalies relative to 1949-1969 mean;

c) Southern Oscillation seasonal pressure index;

d) Rainfall at Ocean Island (0°52'S, 169°35'E), mean monthly anomalies for each season relative to 1950-1969 mean.

Ticks along the top and bottom margins denote the northern autumn of the year immediately preceding.

ADVANCED LECTURES 1982

THE CLIMATE OF THE WORLD - LECTURE 4

Monthly, seasonal and interannual variability II - a regional and local perspective

by C K Folland

Summary

Climatic variability on space scales ranging from 10^7 km^2 (mainly N Atlantic and North West Europe) to local areas within the UK (10 km^2) is discussed with emphasis on relations between the characteristics of variability in space and time. The point is stressed that a comprehensive study of monthly, seasonal or interannual variability has scarcely commenced so the bulk of our digested knowledge exists mainly on local and regional space scales. Even here, the format of analyses have been dictated often more by practical than theoretical considerations, so our understanding of the causes of variability is, not surprisingly, meagre.

1. Introduction

The current state of knowledge about short period variability has progressed quite modestly since Craddock (1964) wrote:-

"Research work concerned with the slower changes in the atmosphere is at the stage typified by Kepler rather than Newton in that its main task is that of assembling facts and recognising regularities and patterns rather than seeking explanations".

..... perhaps emphasised by:-

"... the characteristic timescales of climate and circulation variability and the preferred (timescales) for spatial coherence (of such variability) have received little attention".

Hastenrath and Kaczmarczyk, 1981.

2. Structure of the subject

Many empirical studies have been made of short period (a month to a few years) climatic fluctuations but the results are difficult to assess and to integrate. No unifying theory of regional interannual variability or variability from month to month, except that on the annual time scale, exists. The analytical procedures vary greatly, geographical scope is usually limited and the purposes of analyses often reflect sharply defined practical requirements. It is above all difficult to gain at present an appreciation of the spatial characteristics of time sequences of short period fluctuations. This reflects the difficulty of designing appropriate analyses though more use has been made of time series analysis in the study of longer term fluctuations (climatic change) (Parker Met O 13 Advanced Lecture 8).

Most studies could be grouped as follows (some combine several forms of analysis):-

(1) Case studies of individual "events".

(2) Multiple case studies: "superposed epoch analysis", "analogue selection" in long-range forecasting (LRF), "multivariate analysis".

(3) Time series studies.

(4) Studies of the frequency or probability of "events".

(5) Studies of associations of one type of data with another (correlation and regression).

(6) Eigenvector analysis

Analyses of types 5 and 6 will not be discussed in detail here but Nicholls (1980) gives a good review of their applications in LRF.

3. What is the problem?

We need to explain or predict variations or "spells" of weather like those illustrated below.

N.B. The word "anomaly" is used to describe deviations of weather elements from their average at a given time of the year; the paper is mainly concerned with N Atlantic and N.W. Europe area "anomalies".

EXAMPLE 1 - JANUARY 1963 (Figures 1A, B, C, D)

January 1963 was persistently "blocked" and very cold as was much of the 3 months December to February, with a surface high north of Scotland. Winds were mainly easterly over UK therefore. The Iceland Low essentially disappeared and baroclinic activity (and the "jet stream") was split between a branch south and another north of the block. January 1963 was the coldest month since 1740 in some areas (Murray 1966, Namias 1963) and was made colder by exceptionally low North Sea temperatures that developed. (See also section 5.2).

EXAMPLE 2 - JANUARY 1974 (Figures 2A, B, C, D)

January 1974 was stormy; the Iceland Low was exceptionally deep and persistent. This mild and westerly month in N.W. Europe was an extreme example of the "spell" of four such winters from 1971-72 to 1974-75 (Wright 1975, Painting 1976).

EXAMPLE 3 SUMMER AND AUTUMN 1976 (Figures 3 and 4)

Summer 1976 was exceptionally dry and warm; the jet stream (Figure 3) was persistently far to the north west or north of UK, and frequently anticyclonically curved in UK longitudes. Consequently summer 1976 was one of the warmest in the last 300 years in parts of England and exceptionally dry (Ratcliffe, Miles, Murray 1977).

The transition to Autumn 1976 was startlingly abrupt and in many ways counter to the normal seasonal changes. The jet stream, from early September, became positioned persistently over or south of southern England (Figure 4). Depressions frequently passed over or were quasi-stationary near this region. Autumn 1976 was the second wettest in many southern areas since at least 1727. (See also section 5.1).

EXAMPLE 4 A YEAR-LONG ANOMALY MAY 1975-AUGUST 1976

Summer 1976 was the culmination of an even more exceptional "spell". The conditions shown in figure 3 were maintained with only short breaks (though the jet stream was generally slightly further south) during May 1975-August 1976 (Ratcliffe, Miles 1977). Consequently May 1975 to August 1976 was the driest 16 month period in England and Wales since at least 1727. A Minister for Drought was appointed as the water supply dwindled and other consequences became severe (Folland 1978 gives summary). Figure 5 shows the mean surface pressure pattern and its anomalies over the period; the positive and negative anomalies near England and Greenland are each exceptional ($\pm 3\sigma$). The 500 mb height showed similarly placed anomalies so that, curiously, 500 mb-1000 mb thickness was generally near average (see also section 5.1). Other exceptional extremely long "spells" affecting UK includes the 1921 drought and the 1878 to 1880 cold spell, both of surprising persistence and intensity.

A general discussion of long "spells" is given in Lamb (1972).

4. A problem of interpretation

Many problems can arise in interpreting the unusualness or extent of weather anomalies from climatic data. The variability of weather elements is difficult to define uniquely, as variability in time and space are intimately related. The problems mainly arise because variations at one point do not have the same amplitude or even phase, if sufficiently distant, as those at others (Rodriguez-Iturbe and Mejia 1974 give a mathematical discussion for rainfall). Figure 6 gives two examples of correlation (r) patterns (a) of monthly summer rainfall at Kew with points over England and Wales and (b) annual temperature near Aachen with points over Europe (somewhat idealised from Schonweise (1979)). Note the difference of scale for a given value of r; monthly rainfall in summer is much more poorly correlated at a given distance than annual temperature. These diagram reflect the fact that the statistical structure of anomalies at points differs from that over large areas. For example the 3σ anomalies in Figure 5, which may be expected to occur about once every 300 years on average (described as a "return period" of 300 years) invite the question -- climatic change? but the perspective alters when it is realised that a 3σ point surface pressure anomaly over a year or so occurs somewhere almost every year. Figure 7 shows a related effect for a short time series - Elmdon in the centre of the Midlands has a higher 5 day mean (7 day binomially weighted running mean applied to the daily means) temperature variance than a series of averaged Midlands temperatures. Unsmoothed daily data would show an even larger effect. Related effects will be mentioned again in Section 8.1 as they have important practical consequences.

5. Two case studies

5.1 1975-6 Drought (May 1975 to August 1976, introduced in section 3 and 4)

Figure 8A shows that the 1975-6 drought covered much of W Europe (Figure 8A actually covers October 1975-July 1976). Ratcliffe (1977) and Ratcliffe and Morris (1976) noted that the 500 mb mean jet stream near UK appeared to move fairly steadily northwards between summer 1974 and summer 1976 (Figure 8B). They considered that the northward movement may have been aided by a steady northwards and westwards recession (at a given season) of sea ice near E Greenland (Figure 8B) which commenced after 1969. The recession was thought to be related to unusually stormy conditions in that region during the "westerly" winters of 1971-2 to 1974-5 (mentioned in section 3). They suggested that the jet stream would tend to be moved

northwards and westwards with the ice edge and the associated strong near surface thermal gradient. Also noted was a sharp decline in Pacific Sea Surface Temperatures (SST) north of 40°W and west of 150°E (figure 9A) relative to that further south in most months from late 1974 to late 1976. The 500 mb-1000 mb thickness gradient in mid Pacific near 45°N was enhanced (at least in association) and 500 mb mid Pacific zonal flow averaged 20% above normal during the drought period (figure 9B). N. Hemisphere summers with noticeably enhanced 500 mb flow and those with weakened flow in this region were compared; on average the former were considerably drier and more anticylonic over England and Wales. Similar effects were noted in Winter, apparently strengthened in the presence of a stronger than normal E. Canadian trough, as observed in winter 1975-6. Figure 10A shows the 500 mb pattern in winter in 1975-6 compared to one composited from winters with strong Pacific flow and a deep E. Canadian trough (figure 10B, in arbitrary units based on "eigenvector" analysis). The generally westerly phase of the quasi-biennial oscillation (QBO) during the drought, at least until Spring 1976, were also considered important in helping to maintain a northerly positioned E Atlantic jet stream (Parker Met O 13 Advanced Lecture No 3). Other factors were also considered (see Ratcliffe 1977). By summer 1976 some of these postulated influencing factors were waning or had disappeared; the persistence and great heat of the drought in the summer of 1976 was ascribed to anomalously efficient conversion of solar radiation to sensible heat, as little evaporation would occur due to the dry ground, and which maintained a strong local thermal ridge (over W. Europe). Finally at the end of the summer, with a decline in solar heating, it was thought that the now altered hemispherical influencing factors forced a sudden change in the position of the jet southwards, and the drought ended very abruptly. These associations or "teleconnections" have yet to be confirmed.

A similar, briefer, drought occurred in August to November 1978 (Warrilow and Bader 1979). This drought also ended abruptly and was followed by a wet, cold winter with the jet stream in a southerly position. Figure 11 shows the jet streams during Autumn 1978, the size of the positive surface pressure anomaly, the area it covered and the extent of the drought. The return period of the rainfall deficiency and the severe effect on the volume of two reservoirs in SW England is shown in figure 12. This drought caused much concern to water engineers coming so soon after the 1975-6 drought.

5.2 1962 to 3 winter: a case study involving the test of a hypothesis

Following on from previous work, and e.g. the ideas of Sawyer (1965), Rowntree (1976b) tested a hypothesis that widespread anomalously warm SST in the tropical N. Atlantic helped to maintain the blocked, cold 1962-3 winter over UK. Work using numerical models had suggested, Rowntree (1972), that such conditions in the tropical N Pacific affected the wintertime mid latitude Pacific while in idealised numerical experiments on "tropical forcing" Rowntree (1976a) confirmed the general possibility of such mid latitude effects e.g. as a result of anomalous warmth or coldness in the tropical oceans, though forcing may be largely confined to the winter half year. At this time of the year the mid-latitude westerly circulation at upper levels reaches as far south as latitude 15°N, above the Trade Winds. Rowntree noted that a wide expanse of N. Atlantic between W. Africa and S. America was over 1°C above average in January 1963

(area marked on figure 13A) with a maximum anomaly of 2.8°C close to the W African coast near 17°N. These anomalies had formed in Autumn 1962, before the winter circulation developed. The results of three pairs of experiments with and without the anomaly are shown in figures 13A, B, C. Experiment 1 started from an isothermal atmosphere while experiments 2 and 3 started from real data, 29 December 1962, and differ only in some model details. The results differ markedly, though experiments 2 and 3 are in broad agreement. All pairs show a maximum fall of pressure centred about 40° (experiment 1) or 25° (experiments 2 and 3) north of the anomaly. Experiments 2 and 3 show marked rises near 60-70°N. A marked easterly wind anomaly with the warm SST is evident near 60°N extending southwards over England in experiments 2 and 3 though the anomalies implied by figure 1A are larger. The fall of pressure well north of the SST anomaly and the rise further north still could be regarded as a forced standing wave in the meridional direction; model results (and the real data) show that this wave-like effect extends high into the tropopshere. Observed surface pressure data for 10 winters associated with high SST at Mindelo, Cape Verde Is (17°N, 25°W) and 10 winters with lower than usual SST at Mindelo were compared to the climatological normal. The average of the warm SST years gives a pattern consistent with that of the 1962-3 winter though the pattern is much weaker (figure 14A). The average of the cold years gives an almost exactly reversed pattern inthe N Atlantic, though again a weak one (figure 14B). This case study perhaps indicates the promise of the complementary use of modelling and observed data, though the results presented in this paragraph require confirmation.

6. Superposed Epoch Analysis (SEA)

The comparison of two sets of 10 winters by Rowntree is a simple example of SEA. Two (or more) separate data classes are typically discriminated according to key patterns (eg positive or negative SST anomaly at Mindelo), or perhaps calendar dates, and the differences between the classes studied, e.g. the geographical distribution of surface pressure. In Rowntree's work SEA was used to support an independently arrived-at hypothesis. More often however the hypothesis becomes the result of the application of SEA (a posteriori hypothesis) The key pattern e.g. as in Ratcliffe's work on the 1975-76 drought, is often identified by trial and error as that best discriminating (by association) betwee the (say) opposite extremes of the data under study. The 500 mb flow in mid Pacifi in one drought study serves as an example. The validity of the results are markedly dependent on the accidents of the data sample as well as the ability t consistently identify the key pattern, unless an independent sample is retained for test. SEA has nevertheless been extensively used in LRF (Nicholls 1980) without provision for an independent sample mainly because of lack of sufficient data. However SEA can be very useful if used with caution. Haurwitz and Brier (1981) provide a balanced, critical, discussion of SEA. A well known example of its use, which has stimulated LRF research, concerns the association of N Atlantic an W European surface pressure anomalies with key SST anomalies near Newfoundland in the previous month. (Ratcliffe and Murray (1970)). Readers are referred to the paper for detail and to Korevaar (1982) who highlights problems of ambiguity in the classification of the "key" patterns in this work.

7. Time series analysis

This technique has perhaps been underused. Meteorological time series are typically "noisy"; nevertheless a simple plot of the data against time is often useful and may be revealing. Figure 15 shows a time series providing unexpected information. Annual maximum values of 2 hour rainfall (Jenkinson (1975) p 31) were calculated for 22 stations in SE England over a 22 year period and the 22 values for a given year averaged (they did not necessarily occur in the same 2 hours). These averaged values are normalised as a percentage of their median value for the 20 year period in figure 15. The time series is surprisingly well correlated from one year to the next for such an apparently transient phenomenon. Rainfall of this type is of practical interest to urban drainage design engineers (see section 8.1).

7.1 More complex analyses are often needed. Power spectrum analysis is one type where variation of a data series in time, e.g. annual Central England Temperature (CET) 1659-1980 (figure 16), is converted to variations of the data according to their frequency or wavelength. Cyclical or "quasi-cyclical" recurrences can be identified; figure 17 shows a power spectrum of annual CET (Mason 1976). Wavelengths near 2.1-2.4 years (QBO?) and 23 and 90 years are prominent. Such spectra can however be misleading; annual CET spectra calculated for independent, shorter periods show substantial variation. A way of studying the variations of data of a particular wavelength (frequency) range in time uses filtering analysis. Figure 18 shows a study of the "23" year filtered series of annual CET where variations near this wavelength have been isolated. The amplitude of the series has clearly varied greatly although the progressive decline with time of its amplitude may not indicate purely "random' behaviour. Also shown is the lack of consistency of phase between such CET fluctuations and the "Hale" or double sunspot cycle often suggested as associated with climatic variability on this wavelength. Figure 19 shows a spectral analysis of daily CET for a five year period 1911-15; the annual cycle has been removed however. This spectrum is typical of daily CET spectra over the last 140 years in that much of the variation occurs on the "LRF" timescale (greater than 10-15 days). The spectrum also highlights a large peak in the variations of CET with a wavelength near 9 months on the extreme left of the diagram. The peak is visible to a lesser extent in many other five year periods and is so large in 1911-15 as to almost dominate the unprocessed time series and clearly requires investigation.

Many other forms of time series analysis are used in meteorology e.g. both Box and Jenkins. See Gray (1976) for references and an application of these techniques.

7.2 It may be more valuable to study the time variation of whole fields of variables. A pioneering study is that of Sawyer (1970) who investigated the time variation of N. Hemisphere 500 mb, 500 mb-1000 mb thickness and surface pressure fields using filtering techniques. He contrasted the spatial behaviour of these parameters on "weather" time scales (less than 10 days) and LRF time scales (in the region of 30 days). He found (figure 20A) that the ratio of variations of 500 mb height on the "30 day" time scales to those on lesser time scales was surprisingly large in some places and spatially very variable. High variance around 30 days is noticeable near UK. He also found

(figure 20B) a generally higher correlation between 500 mb height, 500–1000 mb thickness and surface pressure anomalies on the "30 day" time scale than on the "weather" time scale. This correlation is so high over the oceans that the 30 day anomalies there can be regarded almost as "equivalent barotropic". Recently Wallace in USA has carried out similar analysis (film shown by Lorenc Met-070 Advanced Meteorology

8. Practical Application of Climatic Variation data - frequency analysis

Much of the application of climatic data to engineering and agricultural problems has involved this type of analysis. The numerical values of a climatic variable are ranked and the probability or average frequency of the values, especially very large or very small values, is assessed for planning purposes.

8.1 Long duration rainfall - droughts and prolonged wet spells

Tabony (1977) provides a very detailed description, a statistical model and computer programs (available in Met O 8) of the probability of rainfall amounts in any one month up to any 5 years everywhere in UK. Figure 21 illustrates a typical result and some of the problems concerning the mutual interaction of the spatial and time scales of climatic "events" which have to be considered for practical purposes. (See section 4 for initial discussion). Note first that the pronounced geographical variation in the behaviour of rainfall may be reduced by expressing the rainfall on a given time scale as a percentage of its average value on that time scale. One month "point - ie 'any raingauge'" England and Wales rainfall (EWR) has a markedly frequency distribution different from a series representing EW as a whole. Very rare dry events are drier at any point than events of equal rarity over EW as a whole while very wet events are wetter. The difference is less pronounced over 12 months. This difference of behaviour over different time scales reflects the differences in spatial correlation of rainfall on these time scales; one month rainfall is not so well correlated in distance (it is more variable) as is 12 month rainfall. The fact that point rainfall having a 20 year return period has a similar percentage value (the y axis in figure 21) as the 100 year rainfall over EW as a whole suggests that an ability to transfer water over considerable distances in droughts to very dry places from other places in EW some of which would, almost certainly, be simultaneously wetter would be useful. The ratio of the point to the areal value (usually measured as an absolute rainfall amount) is called the areal reduction factor or, better perhaps, areal adjustment factor (Folland, Warrilow and Kelway 1981).

8.2 Very heavy short duration rainfall

Figure 22 shows the shape of the frequency distributions for several short durations for very heavy rainfall (Jenkinson 1975). This type of information is used in dam, river flood protection (Flood Studies Report Volume I) and sewerage design (shortest durations, see Working Party on the Hydraulic Design of Storm Sewers 1981). Very heavy falls not only appear in figure 22 to become rapidly heavier as they become rarer but they do not appear to tend to a maximum value. The idea of a maximum value is however crucial to modern methods of dam design where safety is the overriding design factor. Maxima are in fact calculated in other ways for this purpose. This problem has recently received attention (Warrilow in discussion of the paper by Folland, Warrilow and Kelway, same volume); perhaps a new aspect of the interpretation of certain kinds of climatic data has appeared. This may affect other short duration

phenomena like wind gusts; it is thought that the problem reflects the difficulties of estimating the rarity of more extreme rapidly moving phenomena that last at a point, ie a fixed observation station, for considerably less time than in the moving frame of reference of the event (eg rain storm).

8.3 An agricultural application

A final, contrasting, application of climatic frequency data is shown in figure 23. Maize, a tropical grass, provides a very good source of winter cattle feed ("silage") but the climate of England and Wales is marginal to its success. Figure 23 shows the number of years in ten that a particularly economic form of silage can be successfully made from maize. Areas where more than two bad years in ten are likely to occur may often not be economically viable for this use of the crop.

8.4 Information about extreme winds is also extensively used in the design of structures on both on land (eg Collingbourne 1978) and at sea eg oil rigs (Shearman 1981).

9. Conclusions

Prospects may now improve for developing concepts about short period climatic variability. Besides the greatly improved capability for integrating numerical atmospheric models, diagnostic data which allow daily atmospheric heat, moisture, momentum and radiation budgets to be continuously monitored are becoming routinely available. These are not only available from the FGGE year experiment (Lorenc Met O 20 Advanced Lectures 1982) but since that year via the Met Office/ECMWF/Reading University Joint Diagnostics Project. The task of improving understanding is very challenging; LRF especially depends for its development on some success. The following words written a long time ago perhaps still apply:

> Our knowledge of the properties of the atmosphere would therefore become of the highest utility to mankind, could the varieties of season, or recurrence of sudden changes of weather, be readily prognosticated. But this investigation is so intimately connected with circumstances which influence different climates, that a general table of prognostics can only be completed by the united labours and experience of the learned in every country.
>
> Robertson (1808)

APPENDIX

Where can I find out more about short period climatic variability and application of knowledge about it in the Met Office?

1. Met O 13 (Synoptic Climatology Branch) - basic research into short time scale and interannual variability, application to long range forecasting and climatic change studies.

2. Met O 3 (Climatological Services Branch) - application to British Standard Codes of Practice, individual construction and engineering projects on land and a sea including some abroad (eg oil industry). Basic research on UK local climatic characteristics.

3. Met O 8 (Agriculture and Hydrometeorology Branch) - Application to crop yield problems and animal mortality problems, including disease, (agriculture) an to the design of dams, storm sewers, river flood protection works, and water supply including schemes abroad (hydrology). Basic research into UK rainfall climatology using combined radar and raingauge data. Investigation of individual extreme events.

4. Met O 20 (Dynamical Climatology Branch) - Studies have started into the sequential nature of long GCM simulations, including simulated time series of rainfall and temperature.

As yet there is no "key" text book giving a clear account of short term climatic variability or its application.

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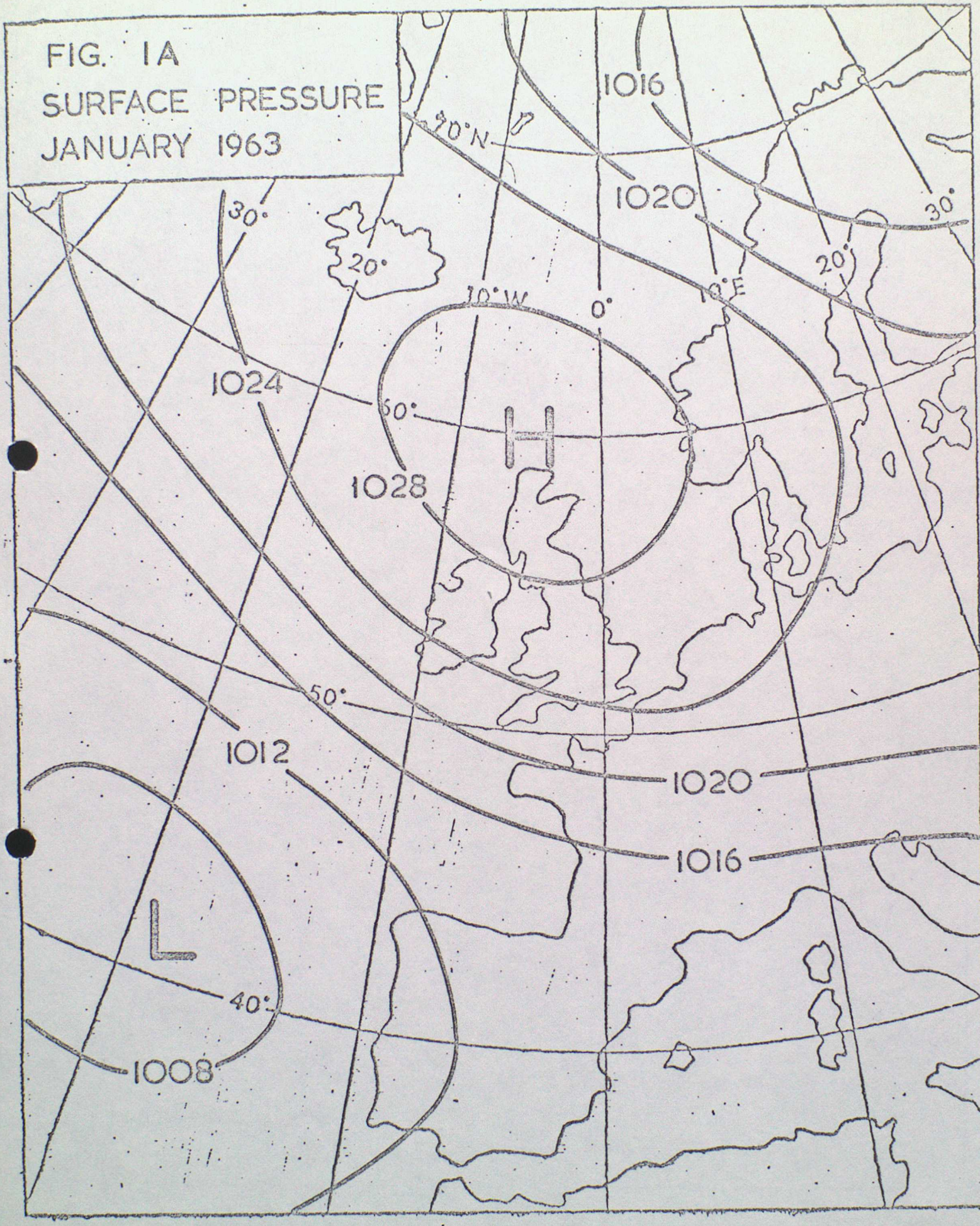
FIG. 1A
SURFACE PRESSURE
JANUARY 1963



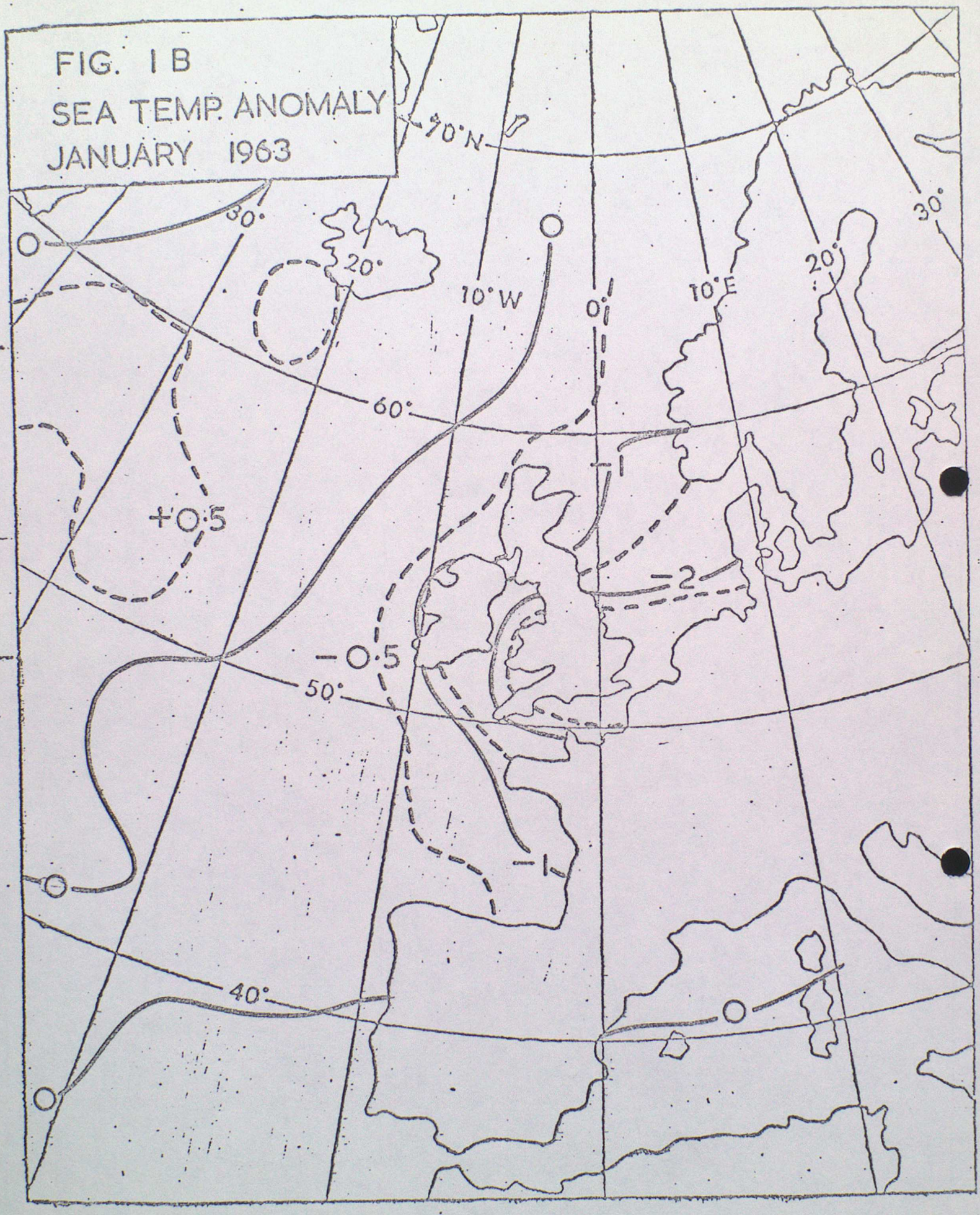
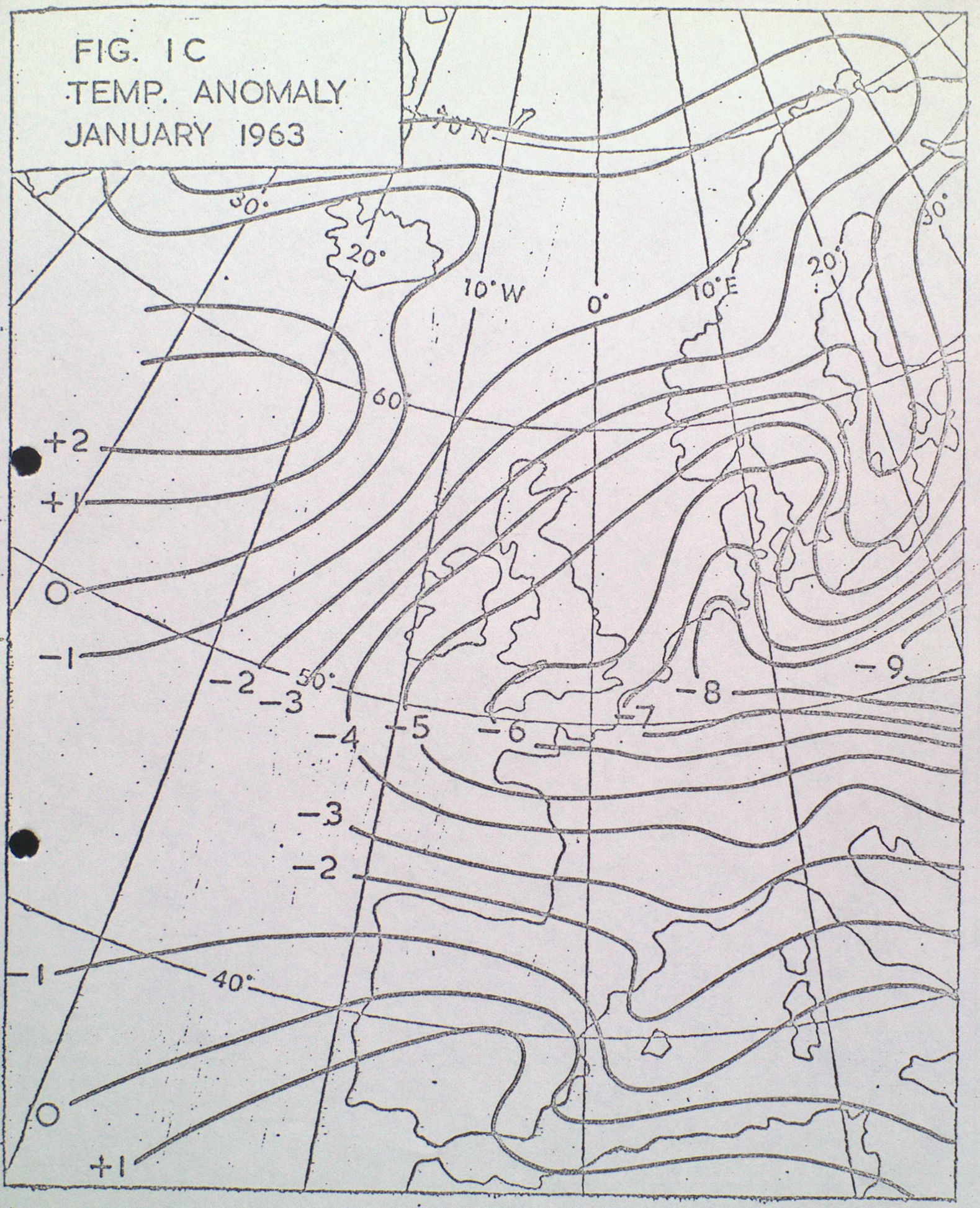


FIG. 1 B
SEA TEMP. ANOMALY
JANUARY 1963
70°N
30°
20°
10°W
0°
10°E
20°
30°
60°
50°
40°
+0·5
-0·5
-1
-2

FIG. 1C
TEMP. ANOMALY
JANUARY 1963



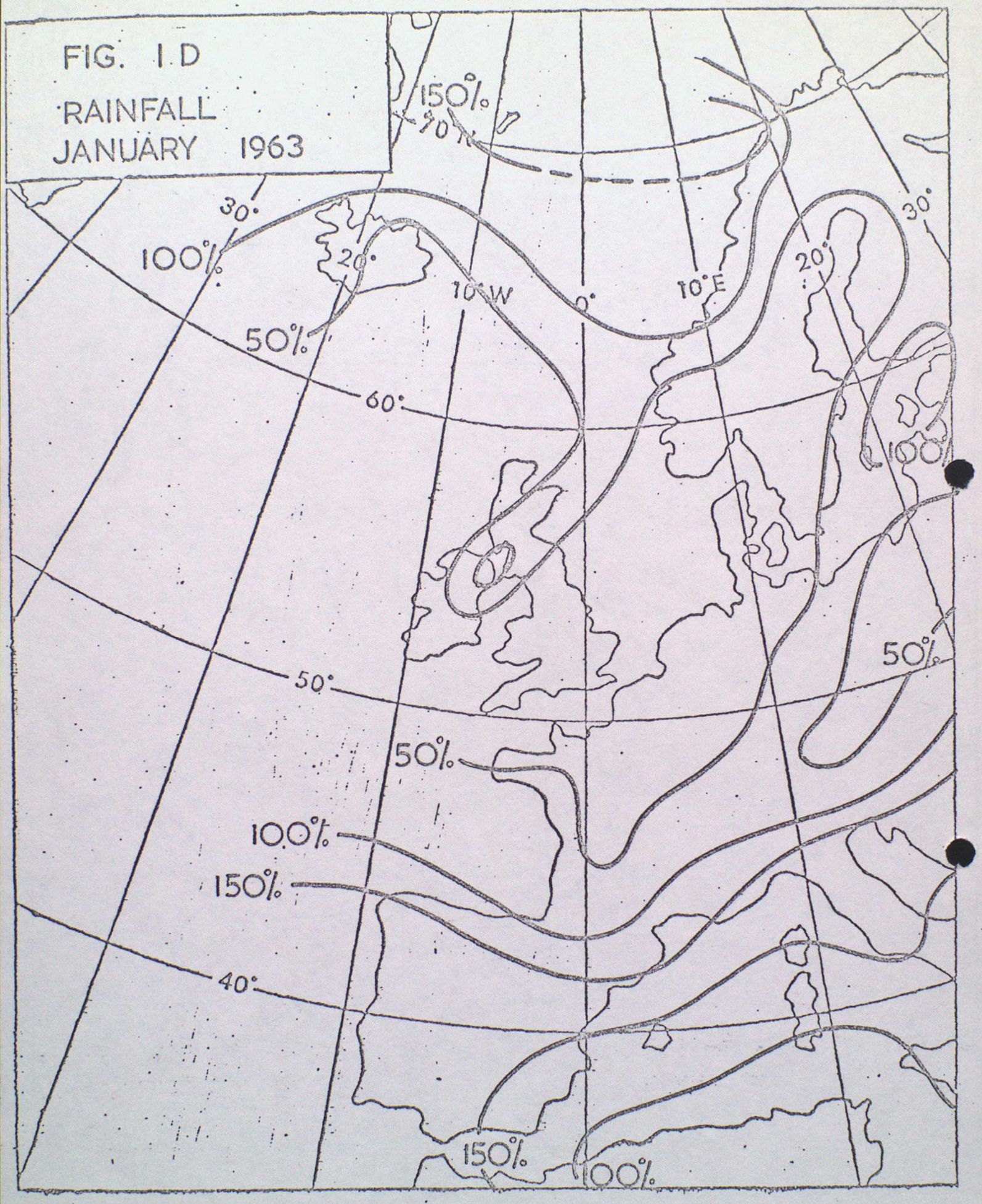


FIG. 1.D
RAINFALL
JANUARY 1963
150%
30°
100%
20°
10°W
0°
10°E
20°
30°
50%
60°
100%
50°
50%
50%
100%
150%
40°
150%
100%

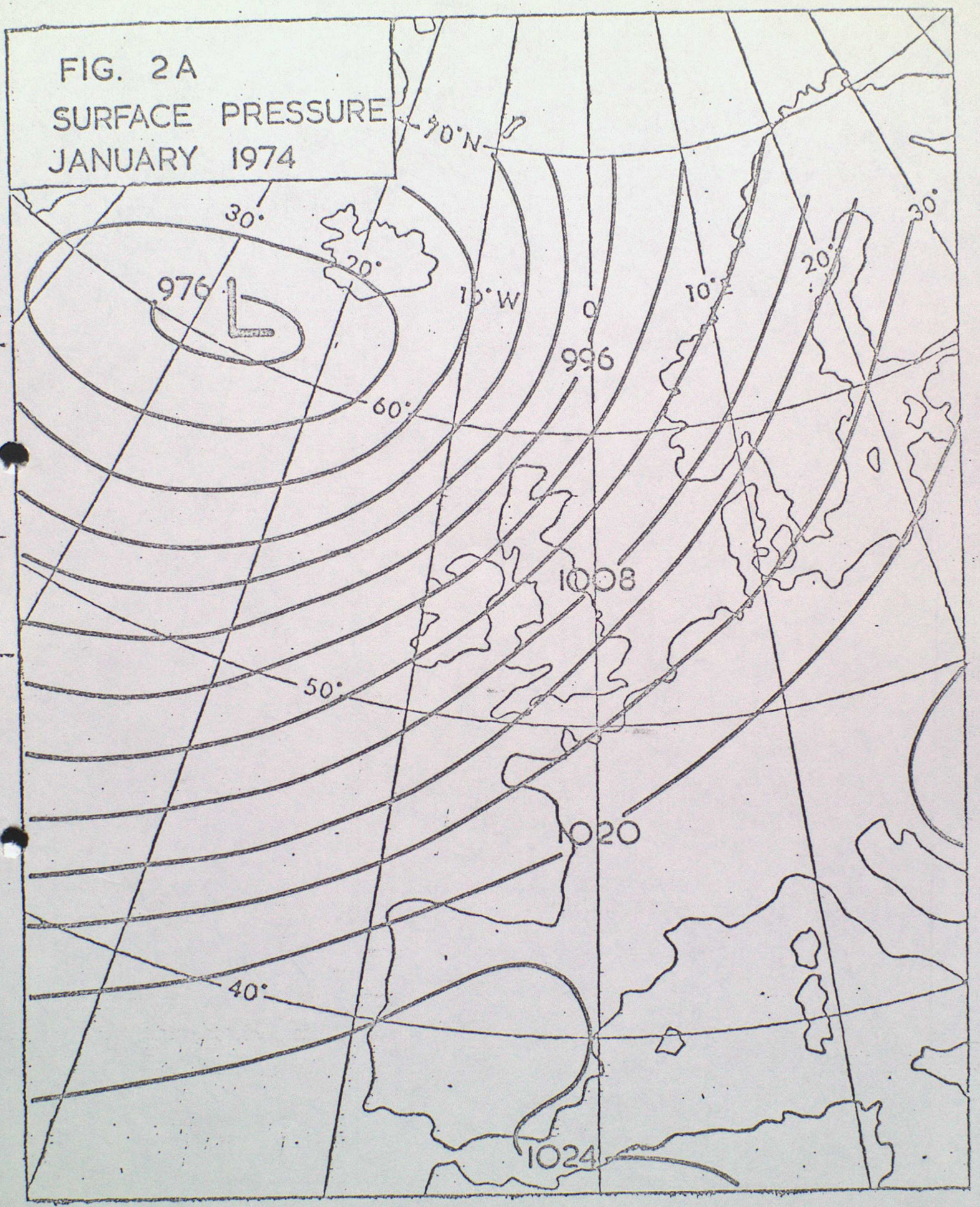


FIG. 2A
SURFACE PRESSURE
JANUARY 1974
70°N
30°
20°
976
15°W
0
10°
20°
30°
996
60°
1008
50°
1020
40°
1024

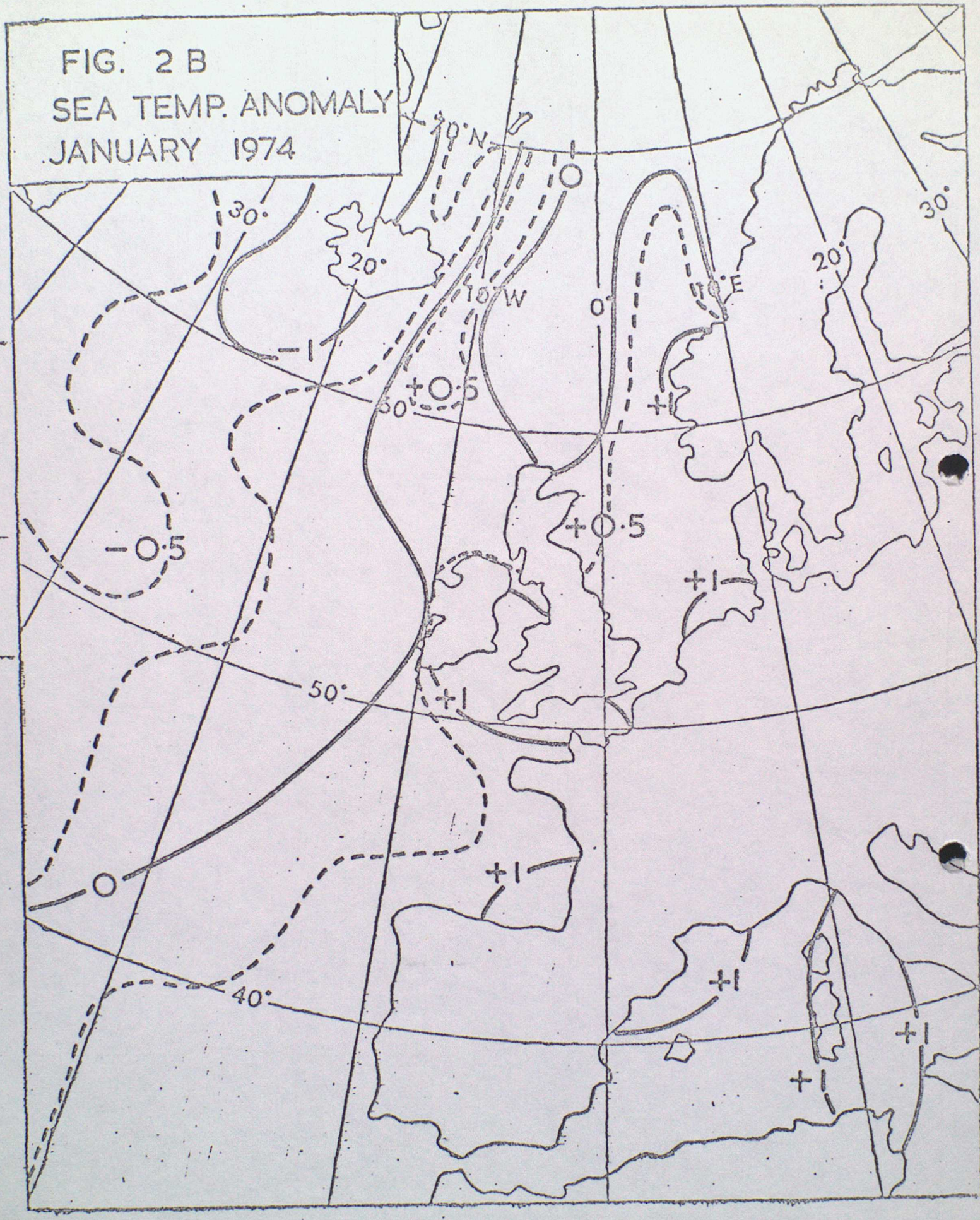


FIG. 2 B
SEA TEMP. ANOMALY
JANUARY 1974

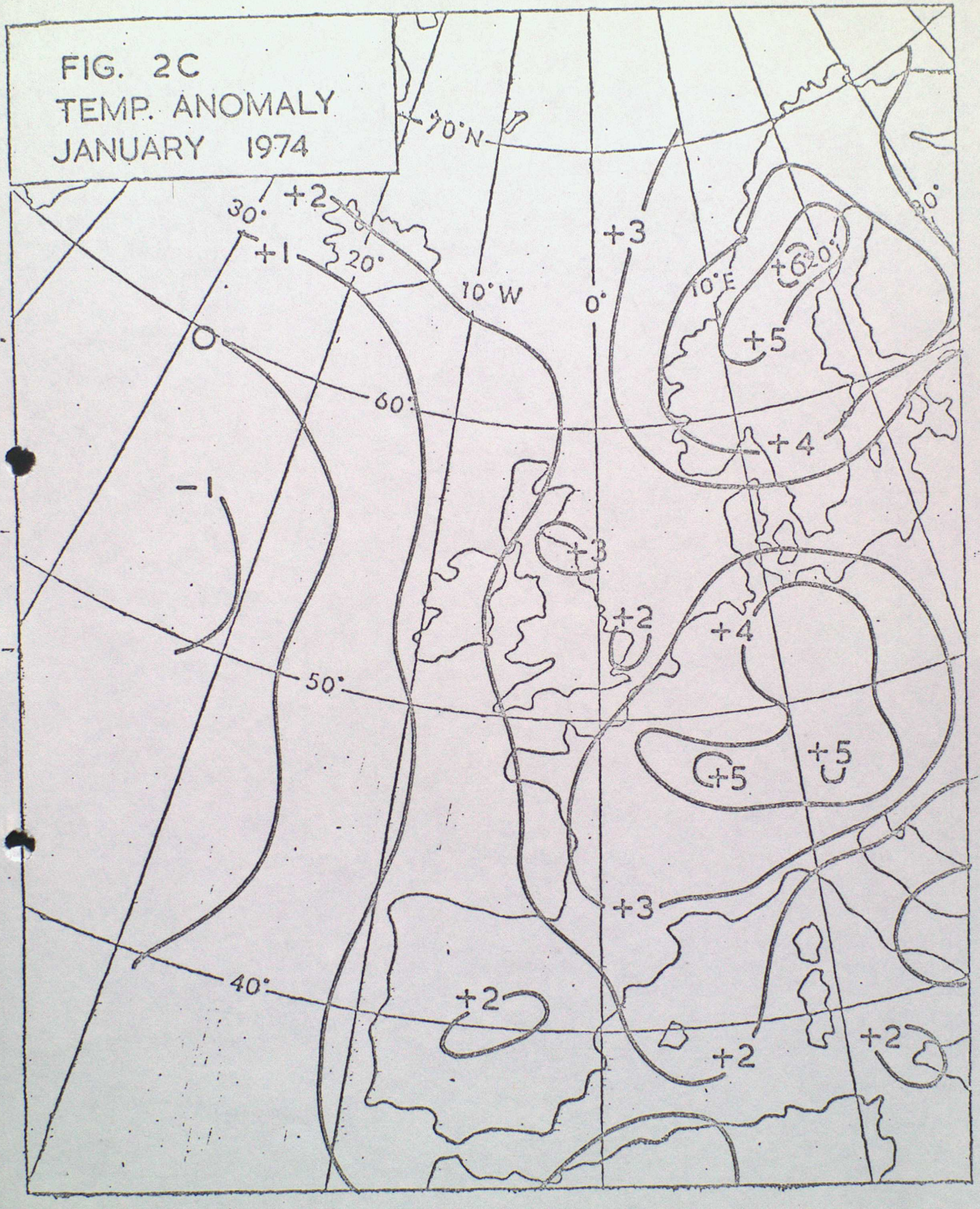


FIG. 2C
TEMP. ANOMALY
JANUARY 1974
70°N
30°
20°
10°W
0°
10°E
20°
60°
50°
40°
+2
+1
+3
+6
+5
+4
+3
+2
+4
+5
+5
+3
+2
+2
+2
-1
0

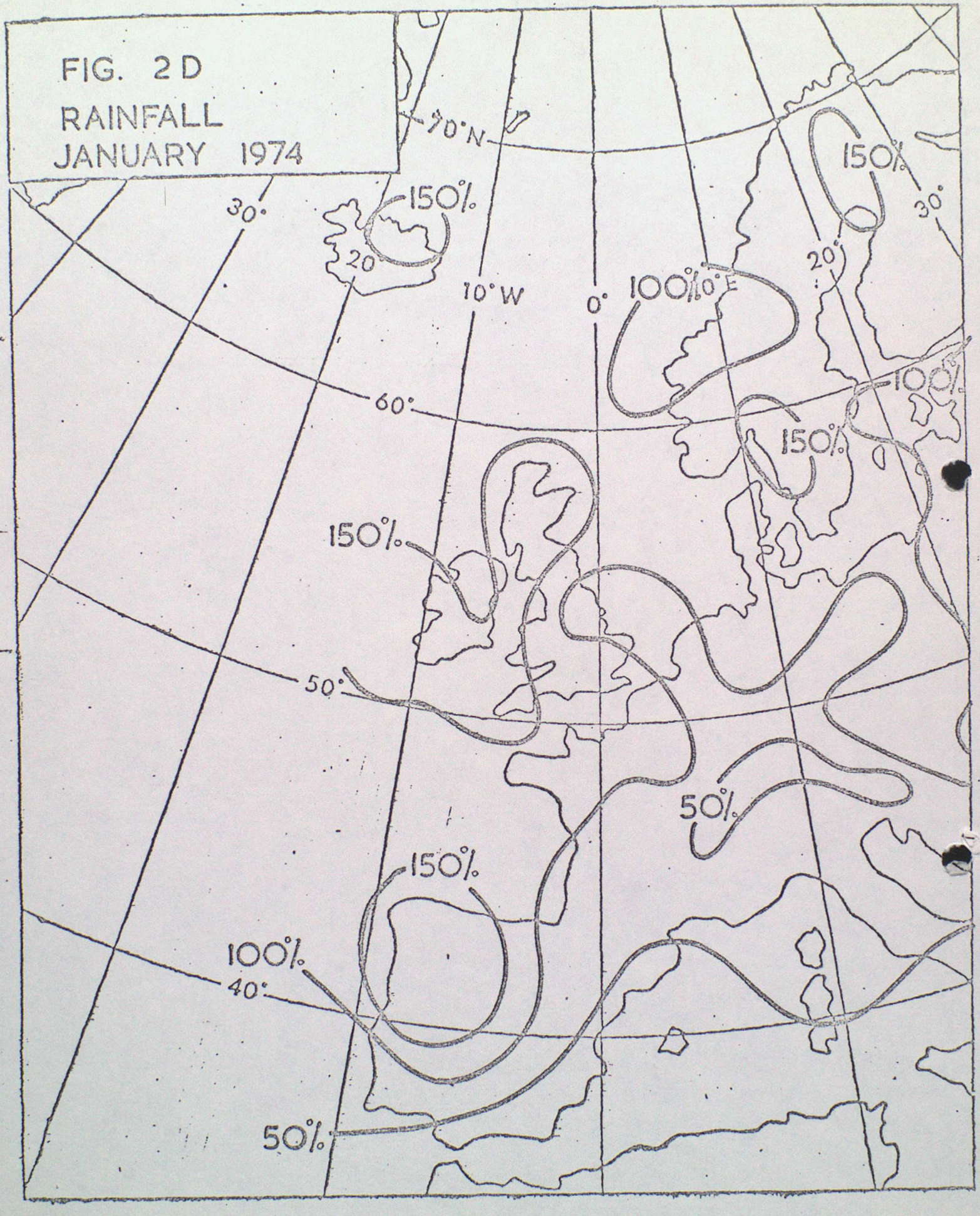


FIG. 2D
RAINFALL
JANUARY 1974
70°N
30°
20°
10°W
0°
10°E
20°
30°
60°
50°
40°
150%
150%
150%
150%
100%
100%
50%
50%

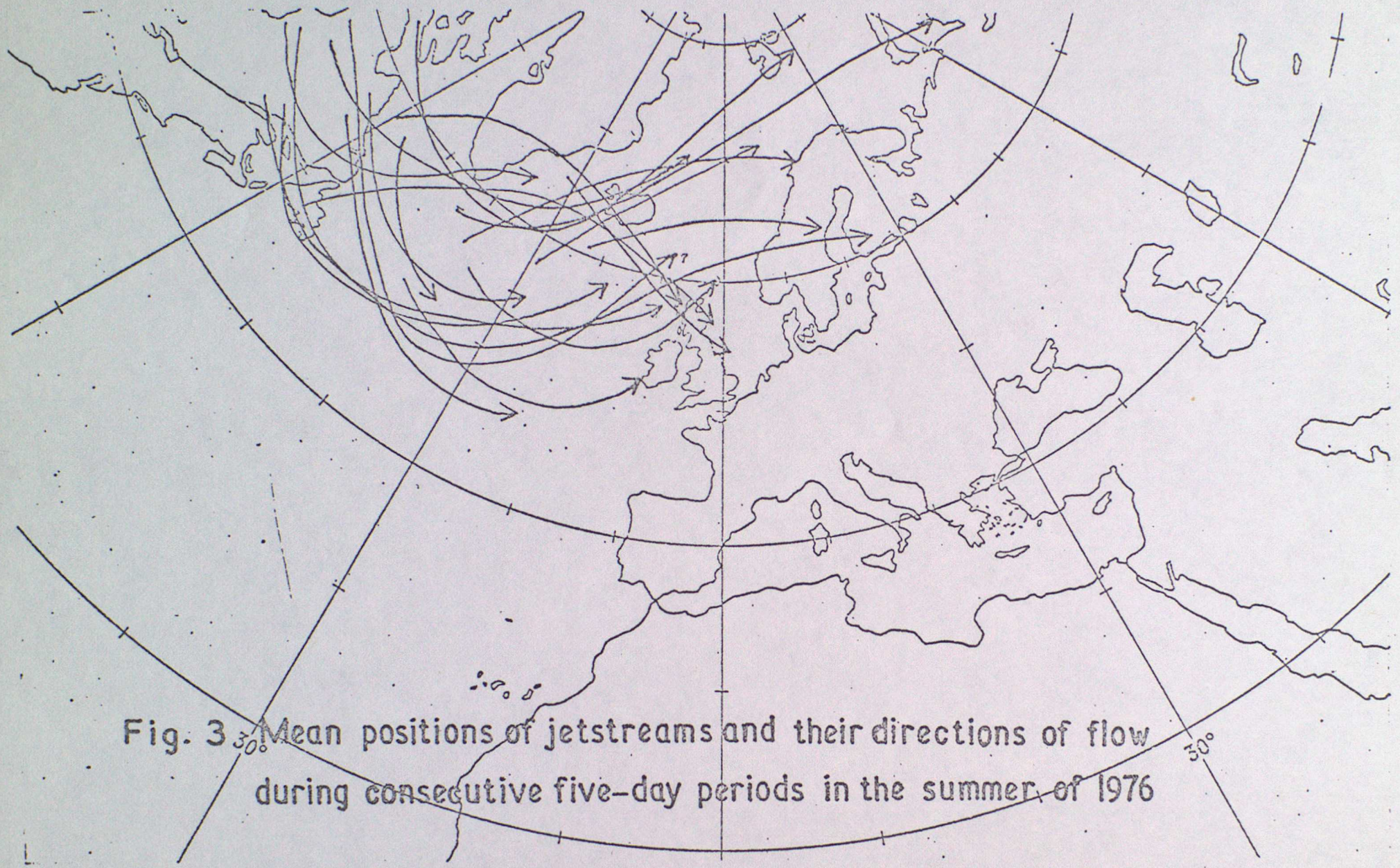


Fig. 3 Mean positions of jetstreams and their directions of flow during consecutive five-day periods in the summer of 1976

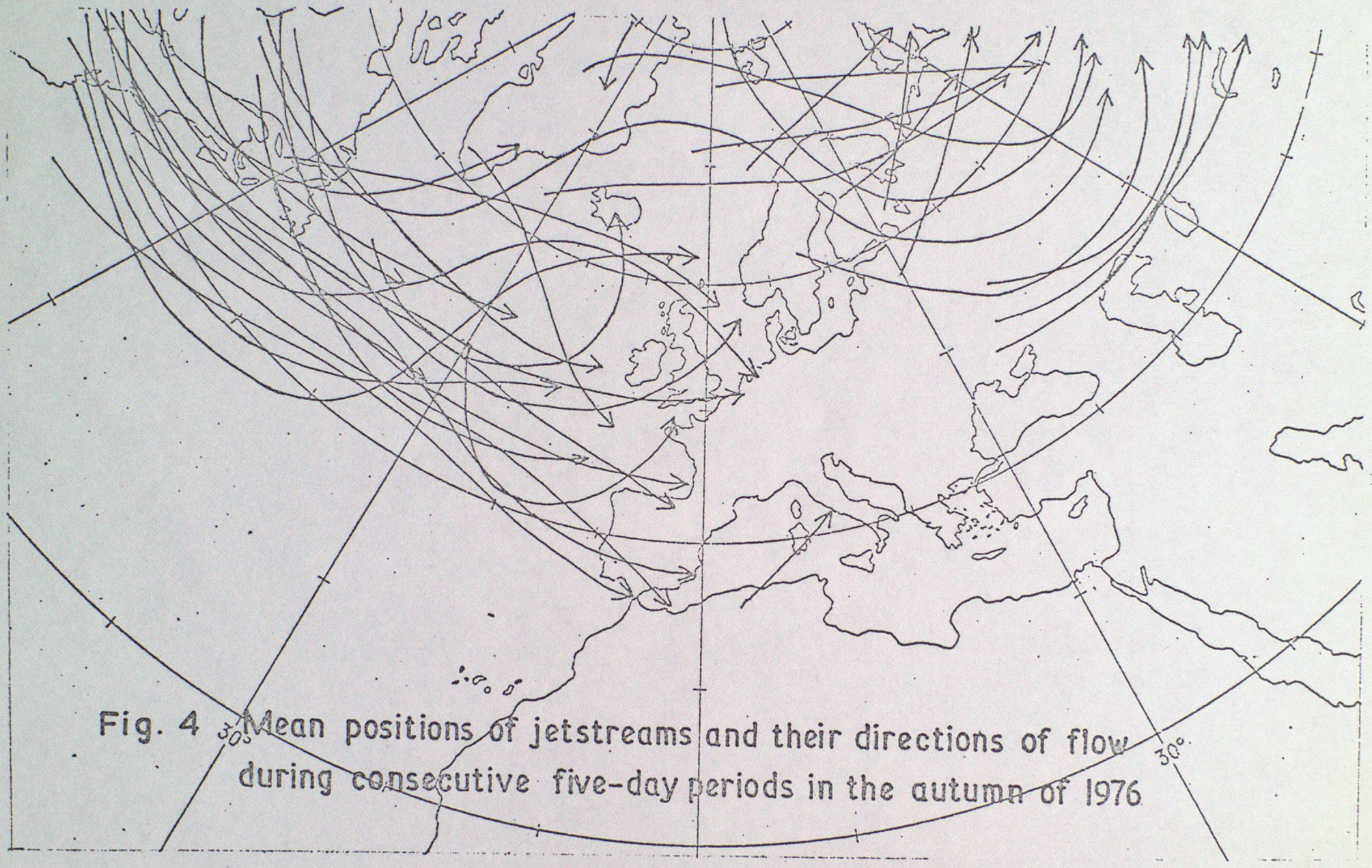
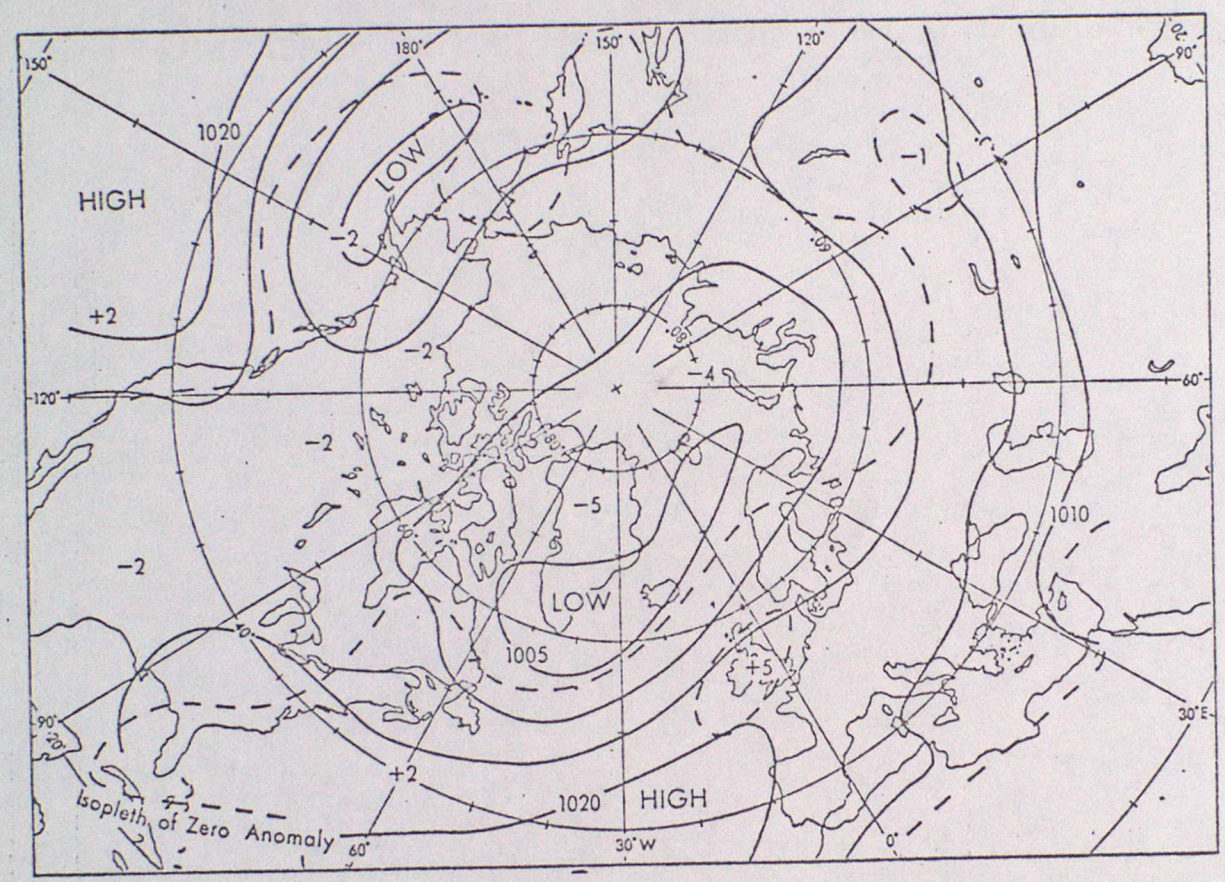


Fig. 4 Mean positions of jetstreams and their directions of flow during consecutive five-day periods in the autumn of 1976

FIG. 5

SURFACE PRESSURE AND ANOMALIES

MAY 1975 – AUG 1976



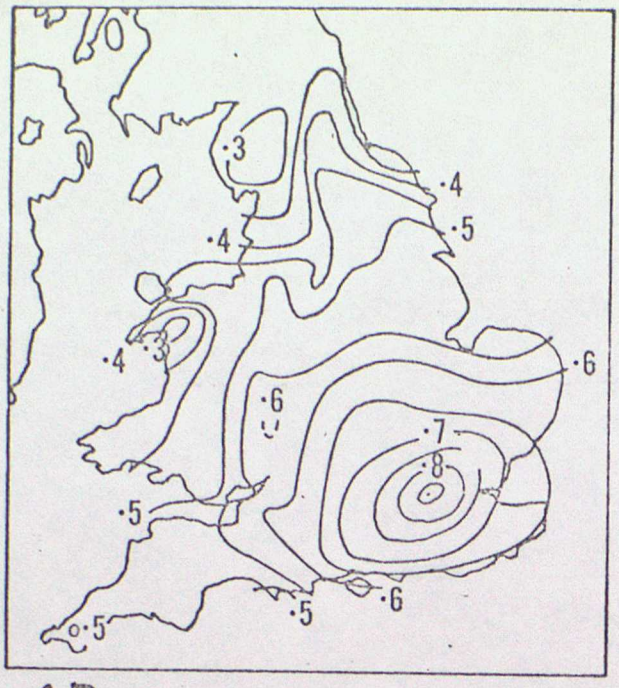
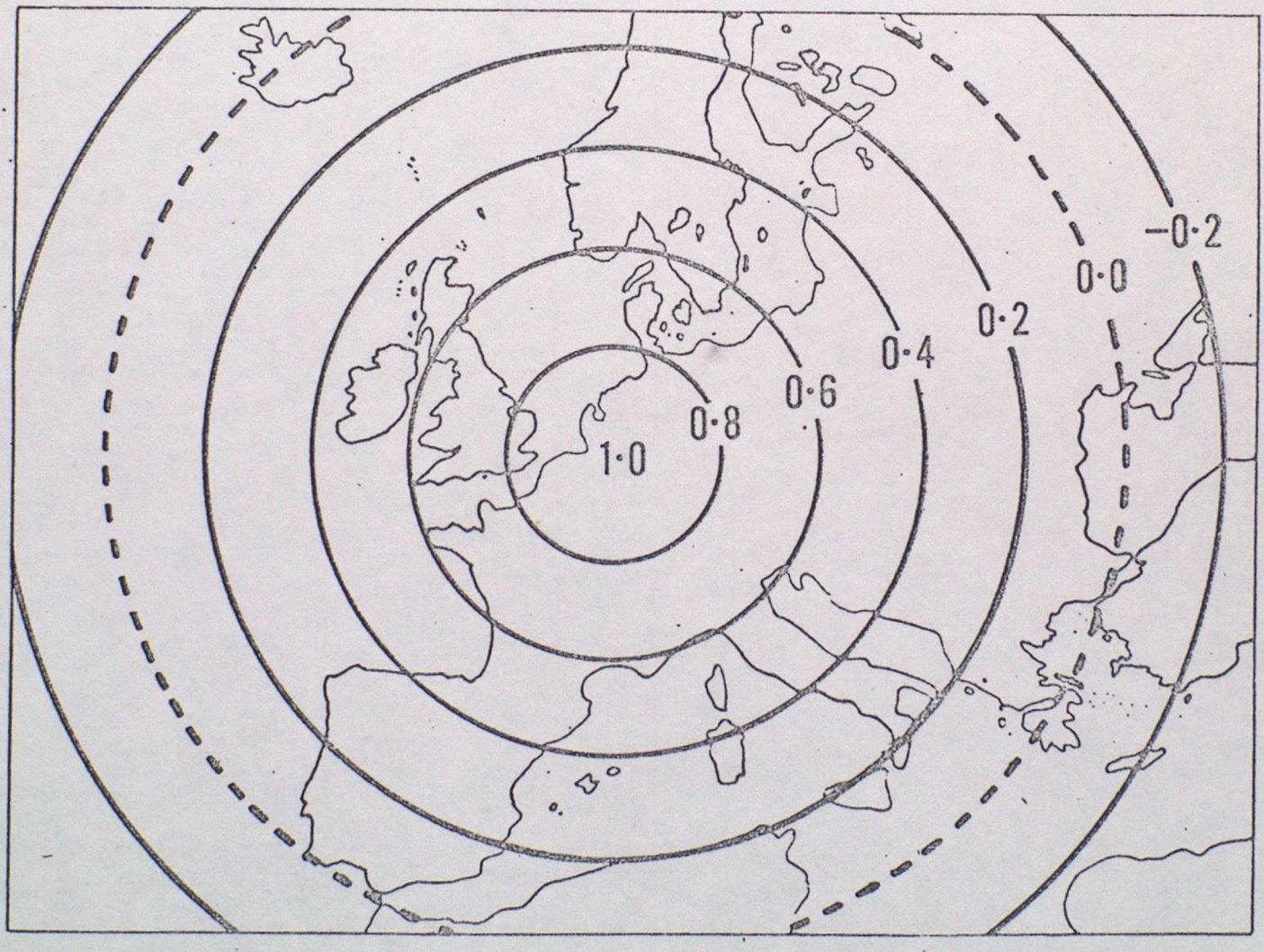


FIG. 6A
Correlation coefficient of monthly rainfall with that at Kew for 1911-70
Mean of June, July and August

FIG. 6B



Simplified Correlation of annual temperature over W.Europe with that near Aachen (1860-1960)

FIG. 7

7 DAY BINOMIAL WEIGHTED RUNNING MEANS

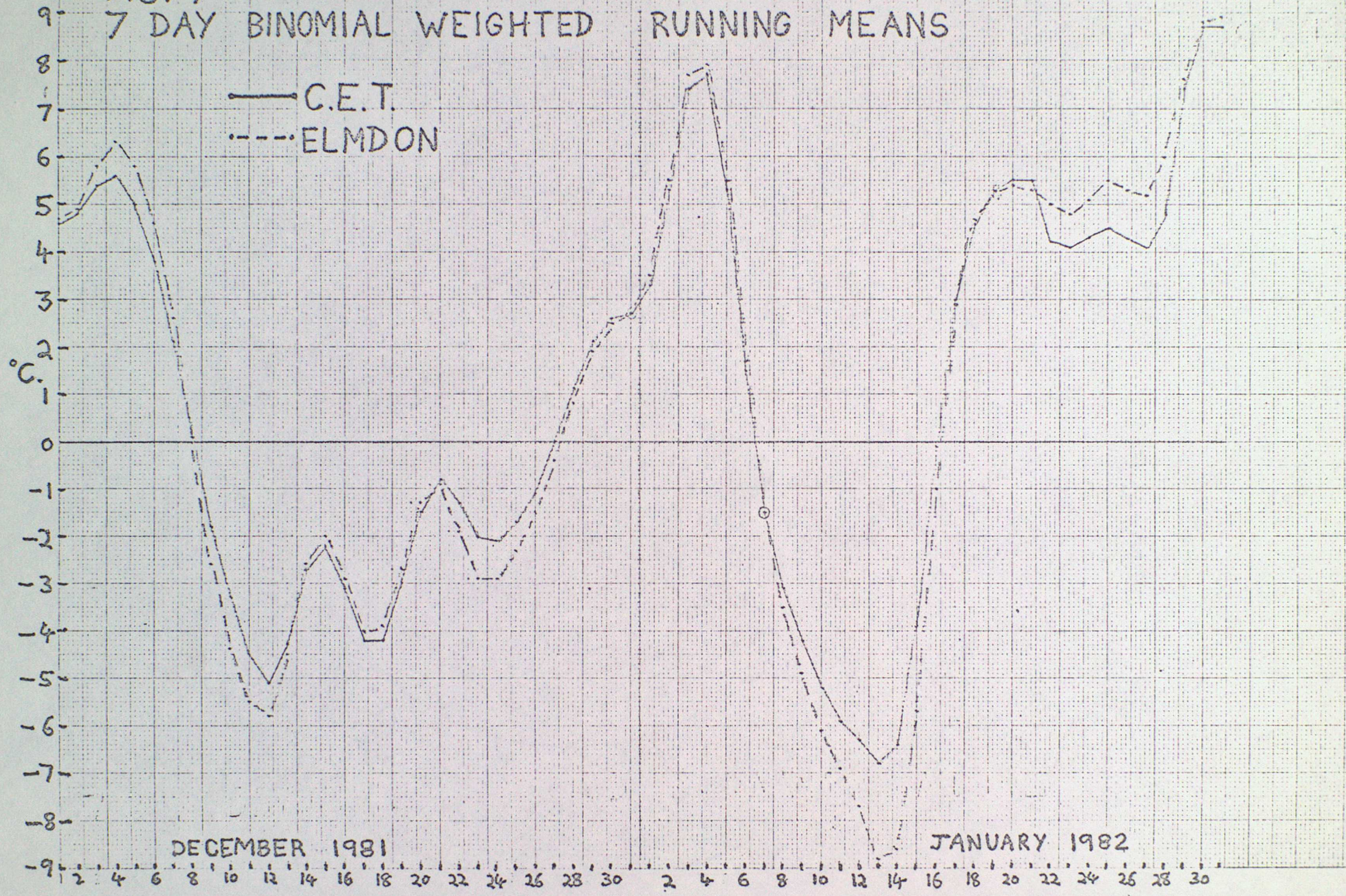


Fig. 8 a

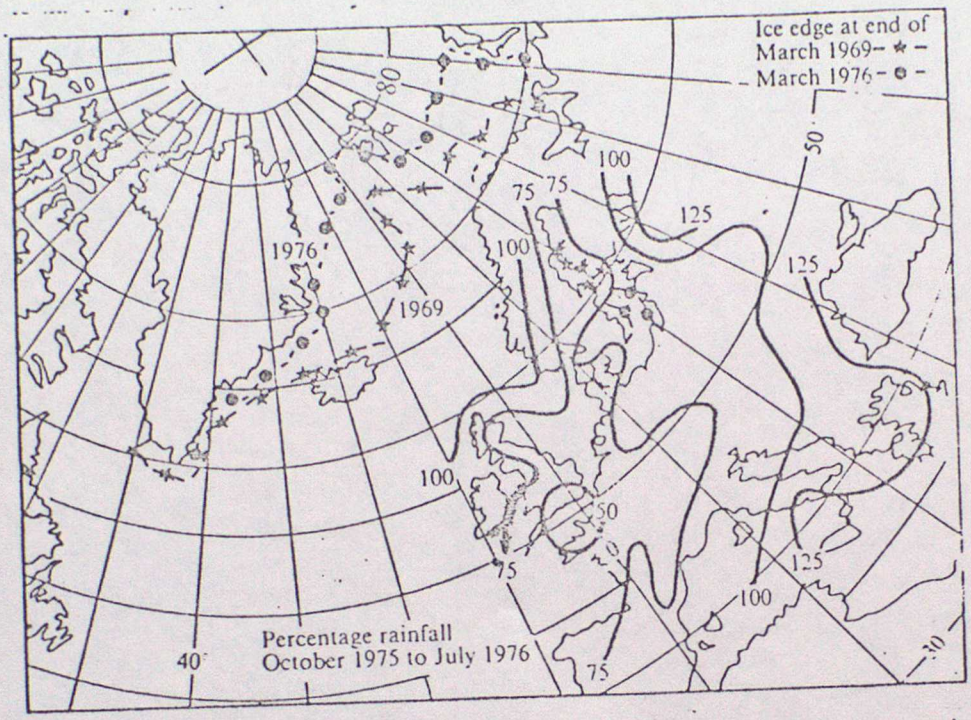

Ice edge at end of
March 1969
March 1976
1976
1969
100
75
125
50
40
Percentage rainfall
October 1975 to July 1976

Fig. 8 b

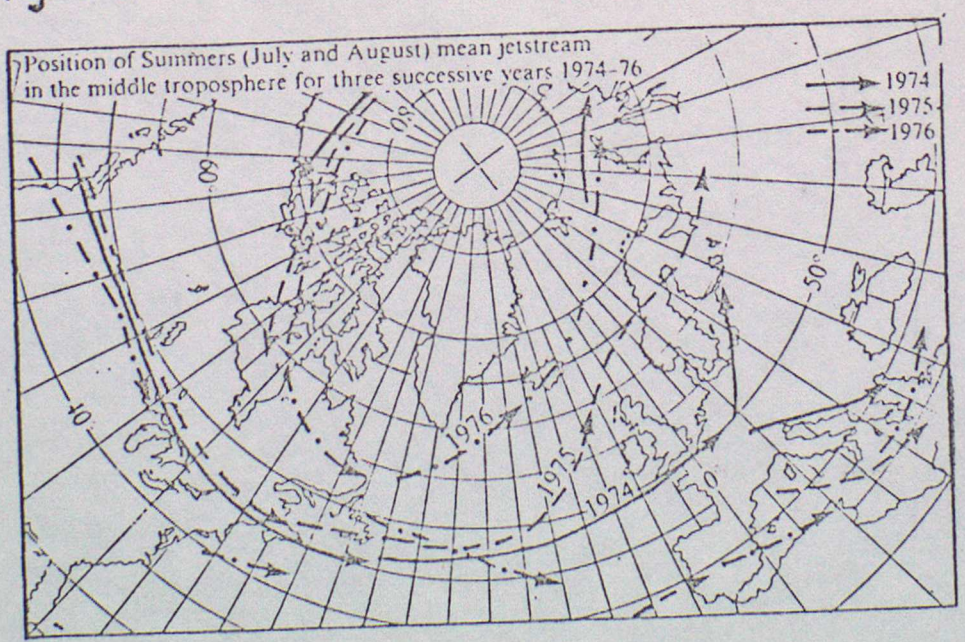

Position of Summers (July and August) mean jetstream
in the middle troposphere for three successive years 1974-76
1974
1975
1976
50
40

FIG. 9A
PACIFIC S.S.T. ANOMALIES
NORTH OF 40°N, EAST OF 150°E

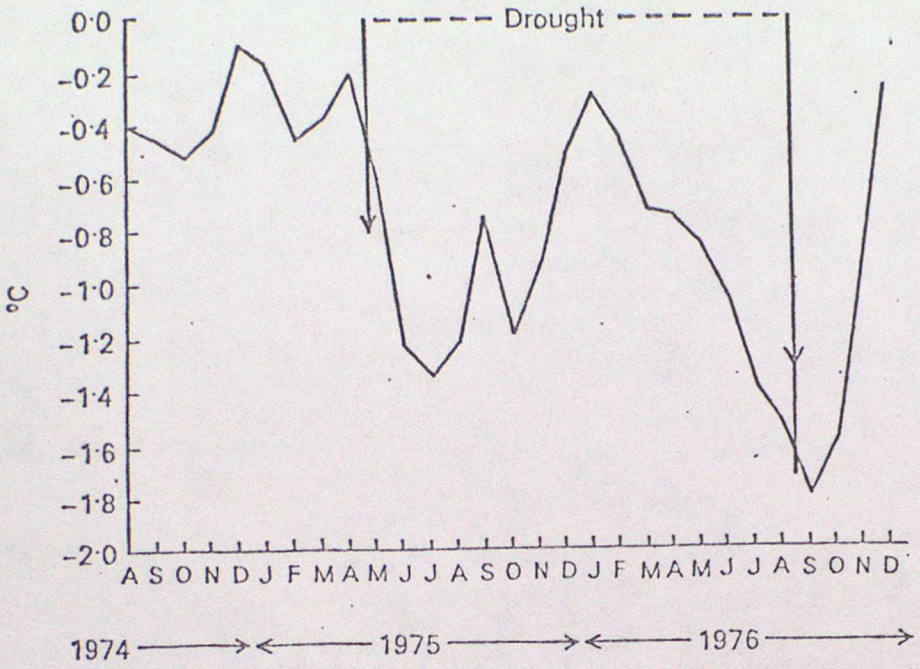


FIG. 9B
500mb ZONAL WIND ANOMALIES
45°N 170°W [Normal 17 ms^{-1}]

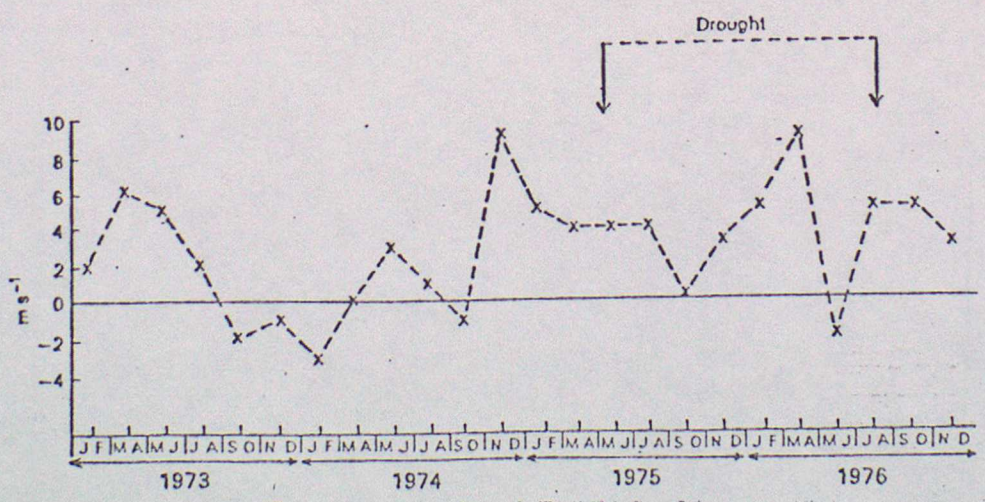


FIG. 10A
COMPOSITE 500mb EIGENVECTOR PATTERNS JANUARY–FEBRUARY
(negative shaded)

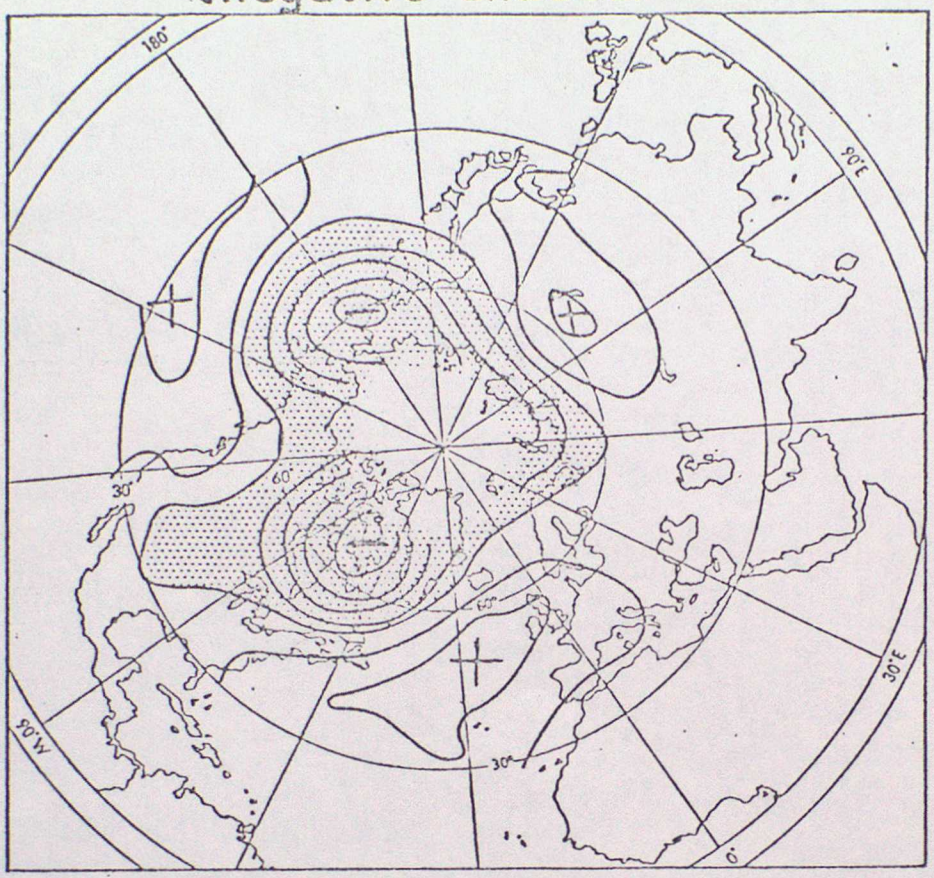
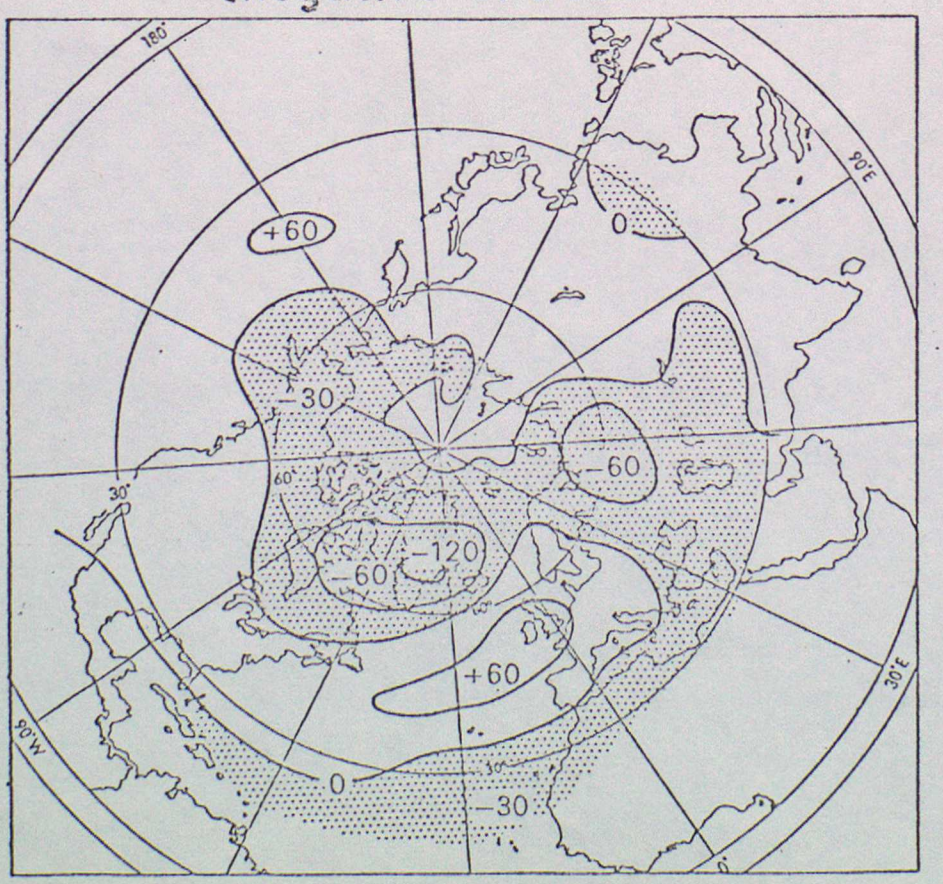


FIG. 10B
500mb ANOMALIES JANUARY–FEBRUARY 1976.
(negative shaded)



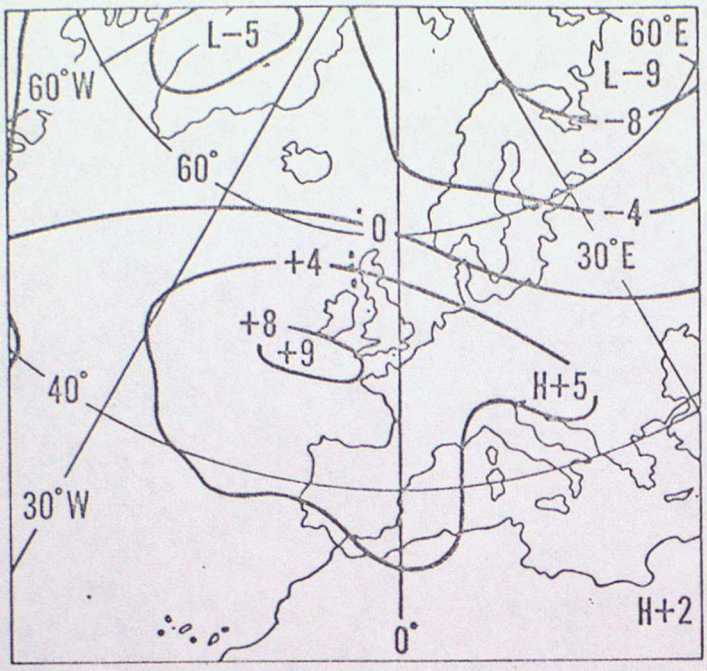


Fig. 11a Mean surface pressure anomaly – September to November 1978

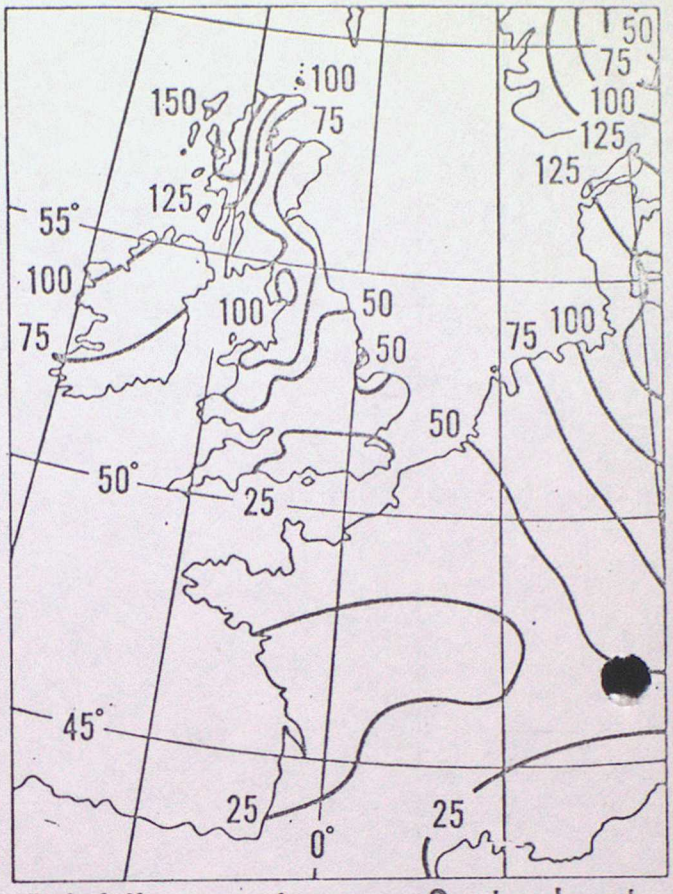


Fig. 11b Rainfall percentages – September to November 1978 as a % of normal

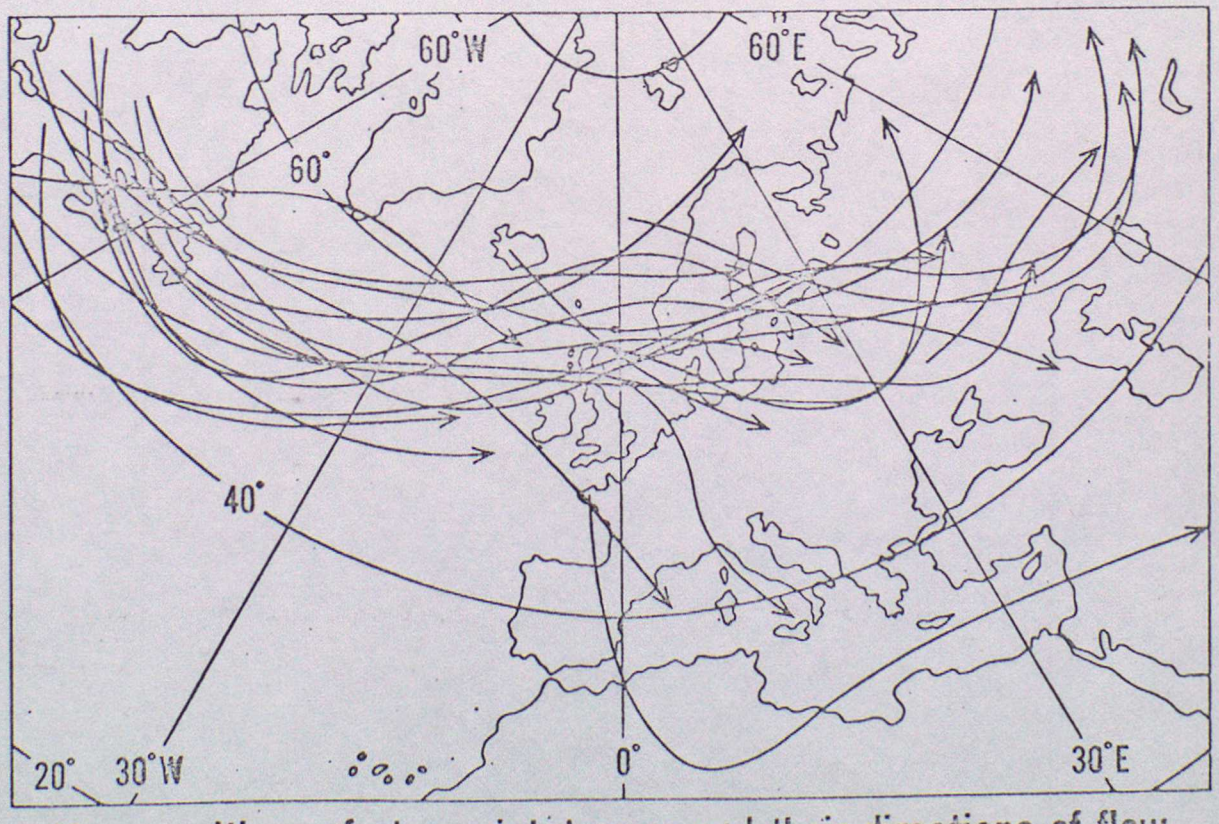
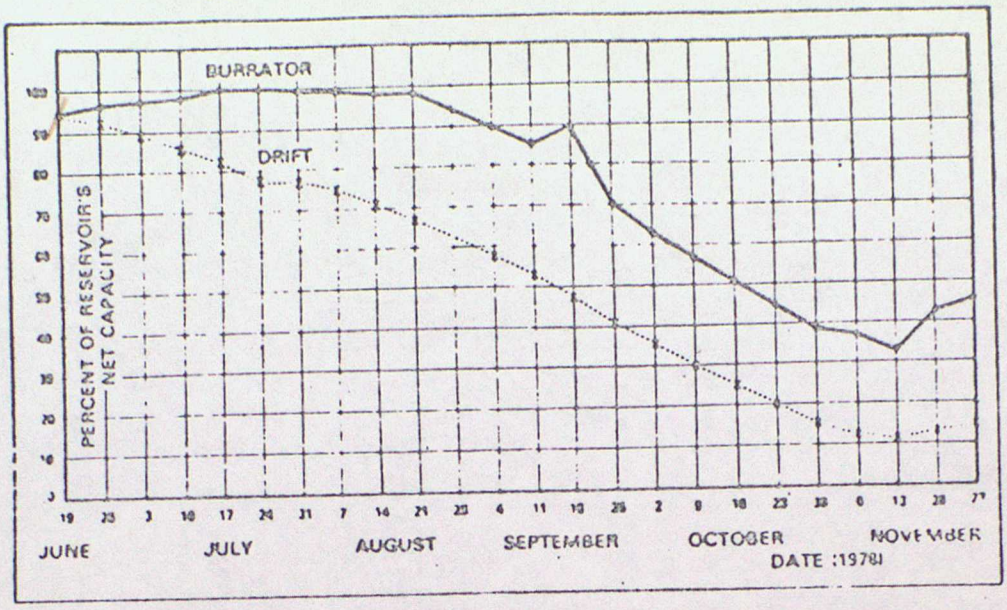


Fig. 11c Mean positions of strong jetstreams and their directions of flow

Fig. 12a

What happened to two reservoirs – Autumn 1978



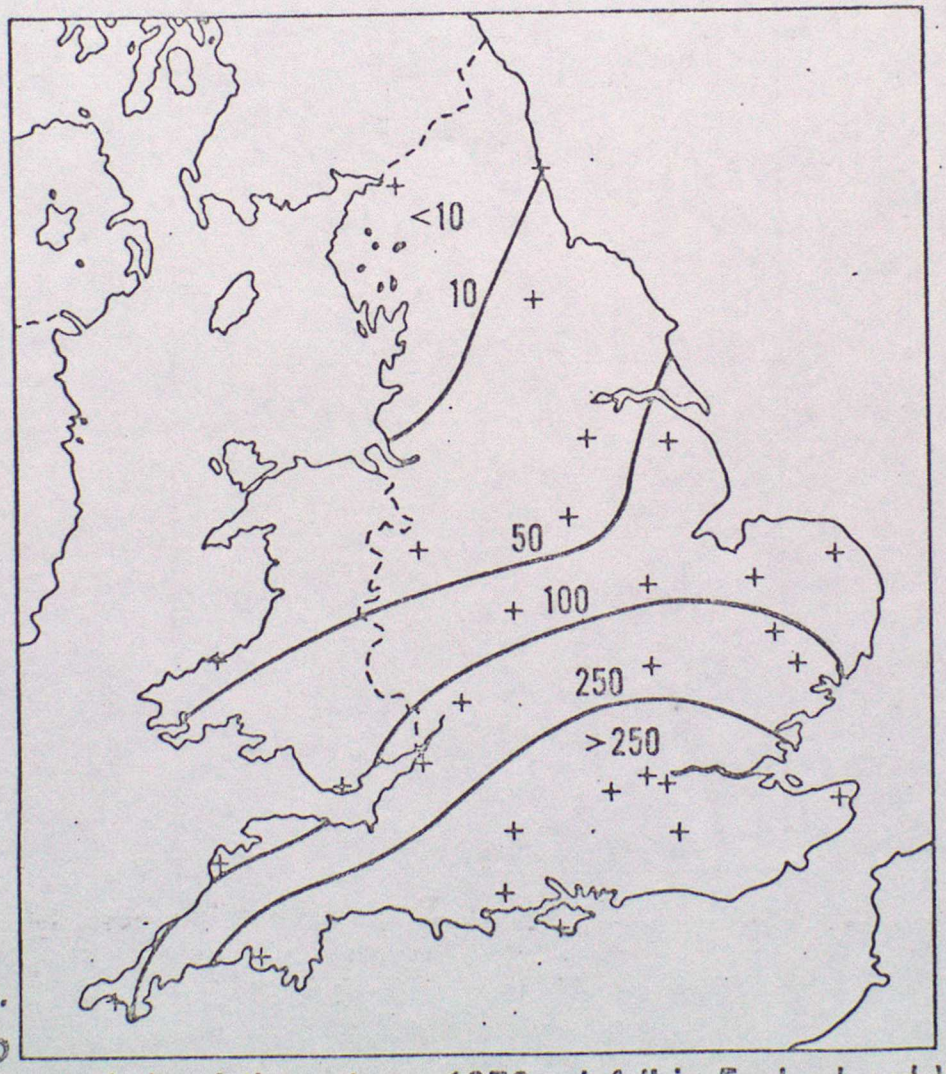
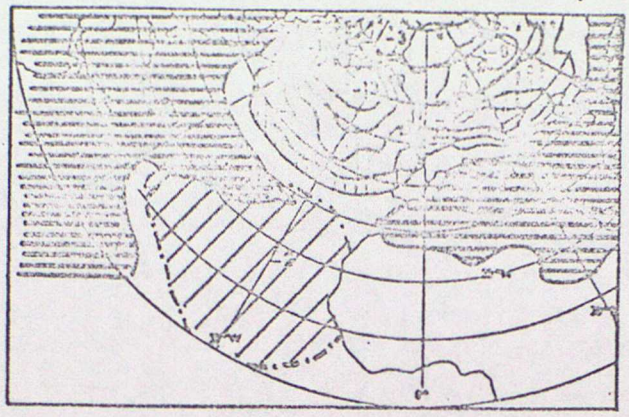


Fig. 12b

Return periods of the autumn 1978 rainfall in England and Wales

WINTER SURFACE PRESSURE ANOMALIES
TROPICAL SST FORCING

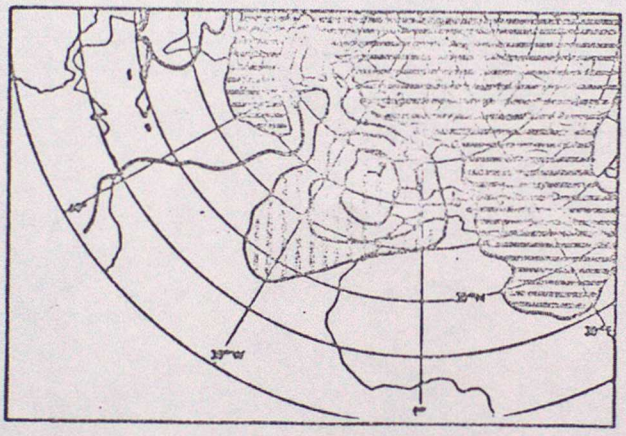
Fig. 13a
INITIAL
DATA
ISOTHERMAL



SST > 1°C.
ABOVE
NORMAL

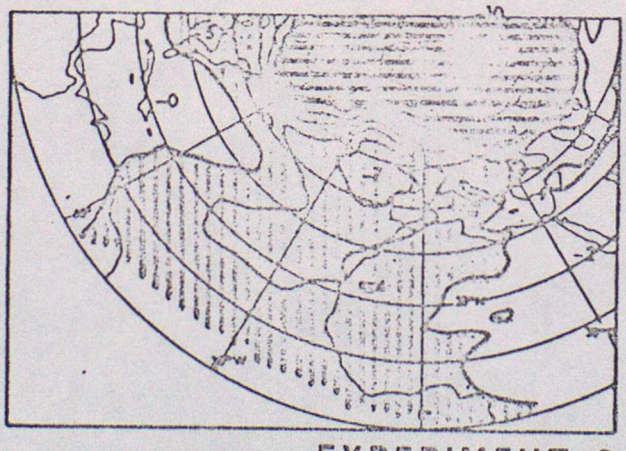
EXPERIMENT 1

Fig. 13b
INITIAL
REAL DATA



EXPERIMENT 2

Fig. 13c
INITIAL
REAL DATA



EXPERIMENT 3

(AFTER ROWNTREE 1976)

FIG. 14A
(10 HIGHEST WINTER S.S.T.) SURFACE PRESSURE, MINDELO, CAPE VERDE IS.

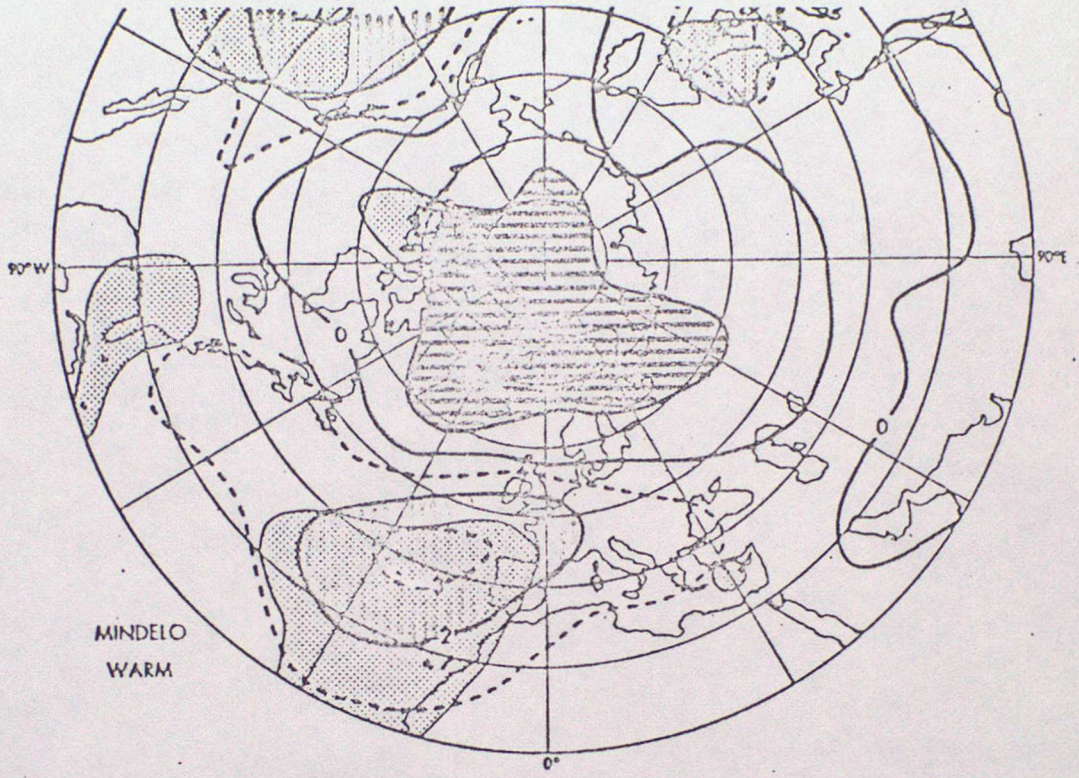
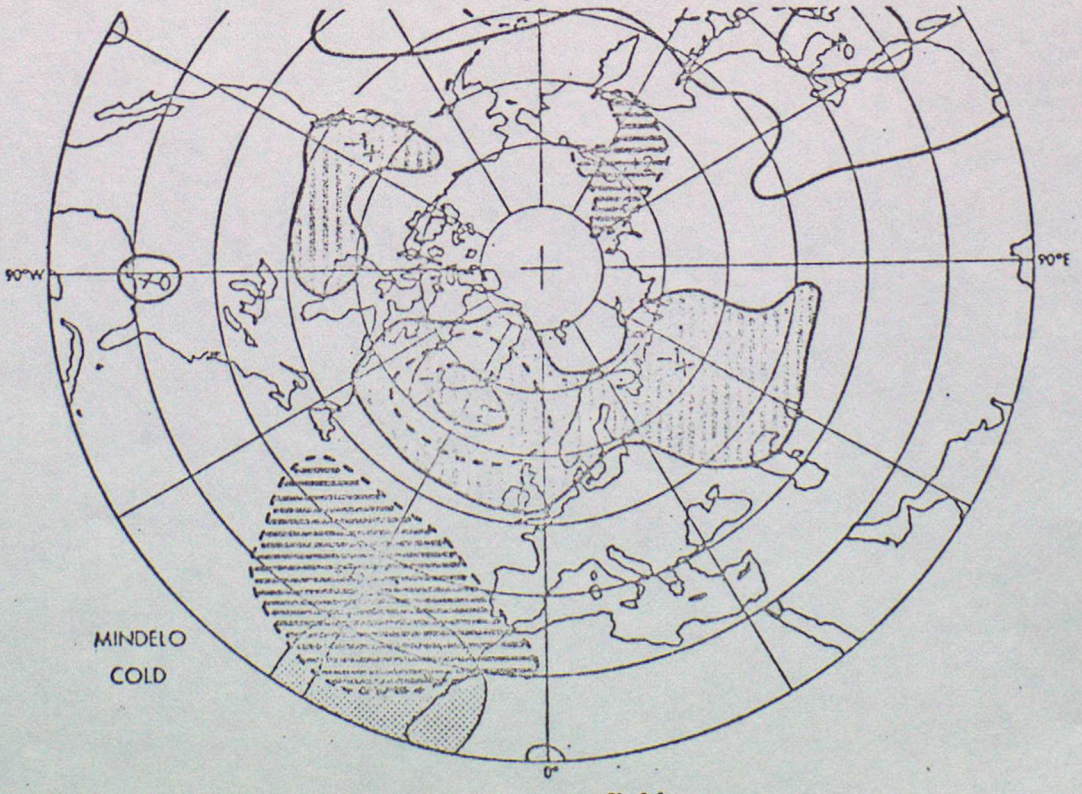


FIG 14B (10 LOWEST WINTER S.S.T.) SURFACE PRESSURE, MINDELO, CAPE VERDE IS.



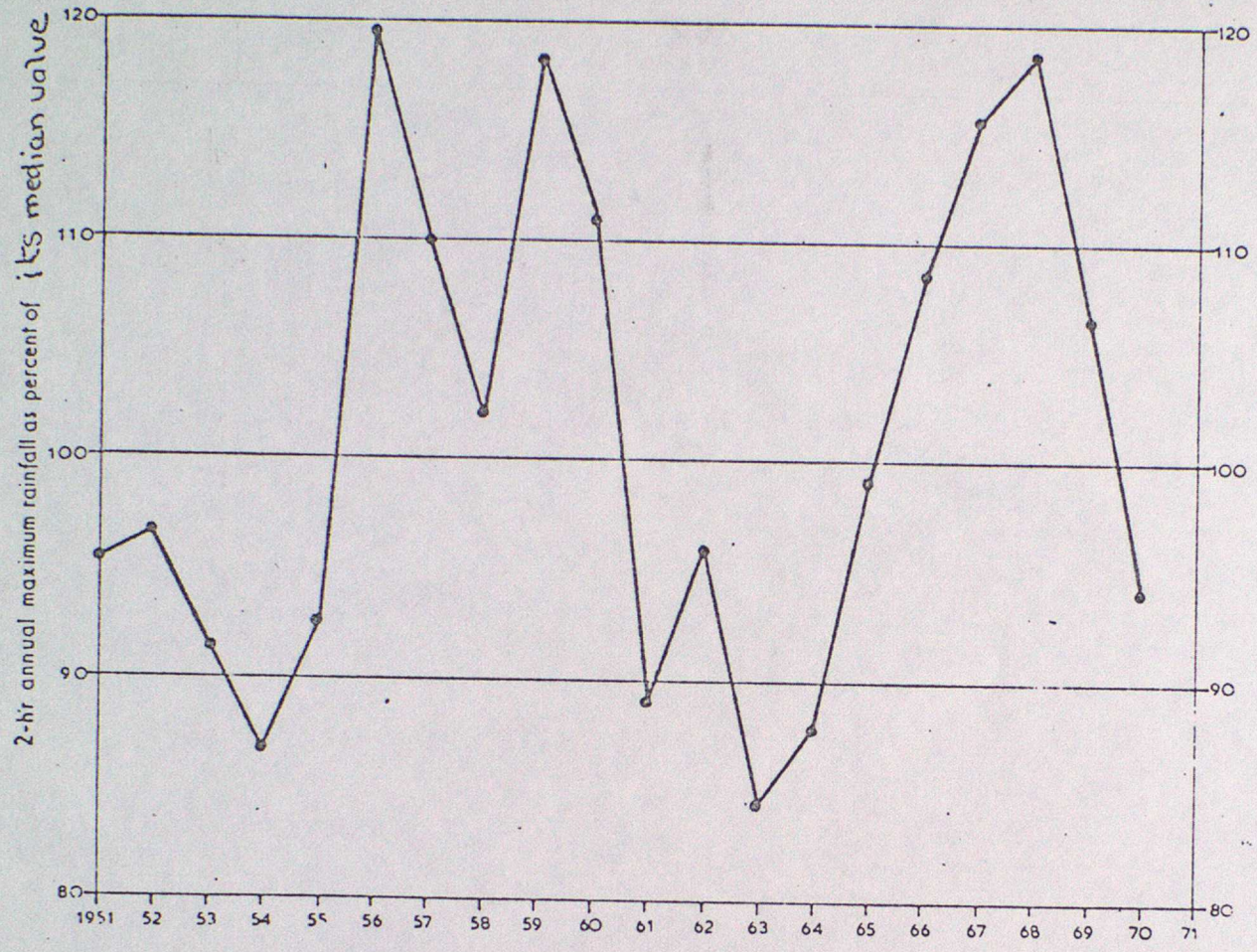
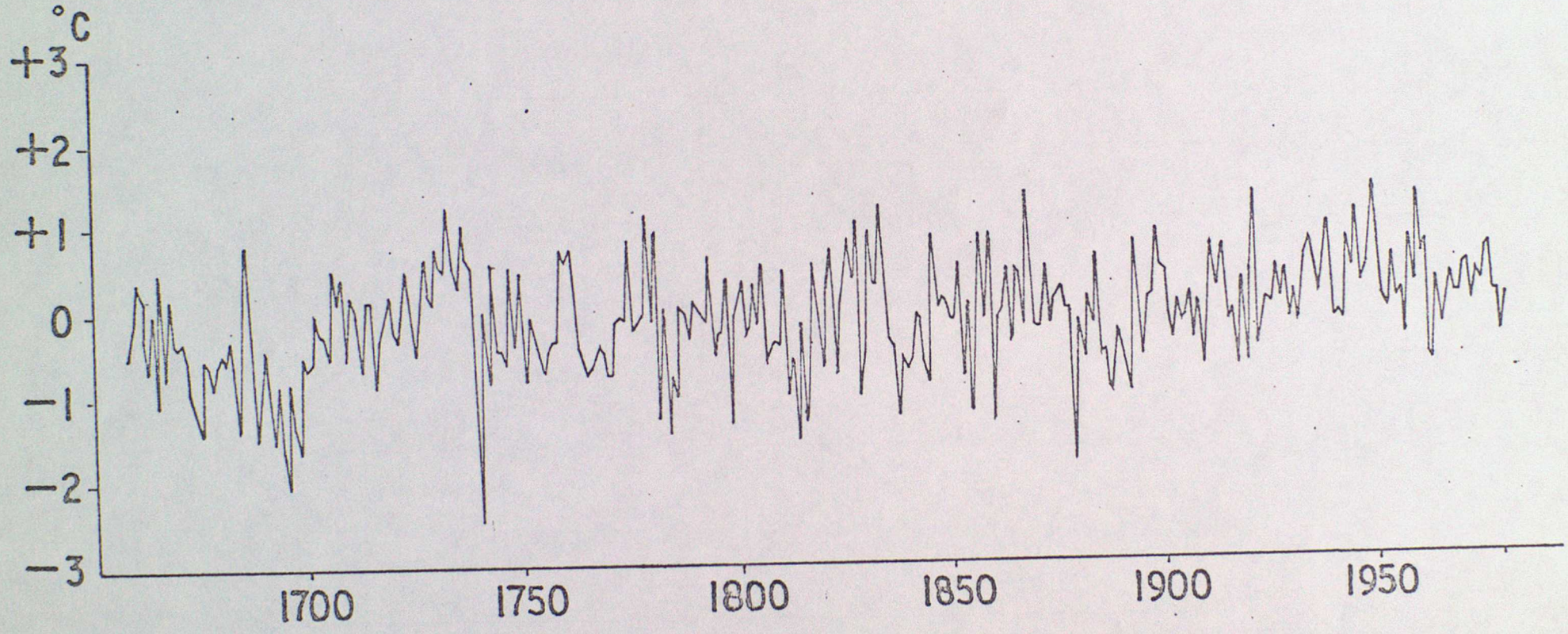


Fig. 15 Variation of 2 hour annual maximum rainfall (median value of 22 stations)

Fig. 16

ANNUAL CET ANOMALIES

1659 – 1980



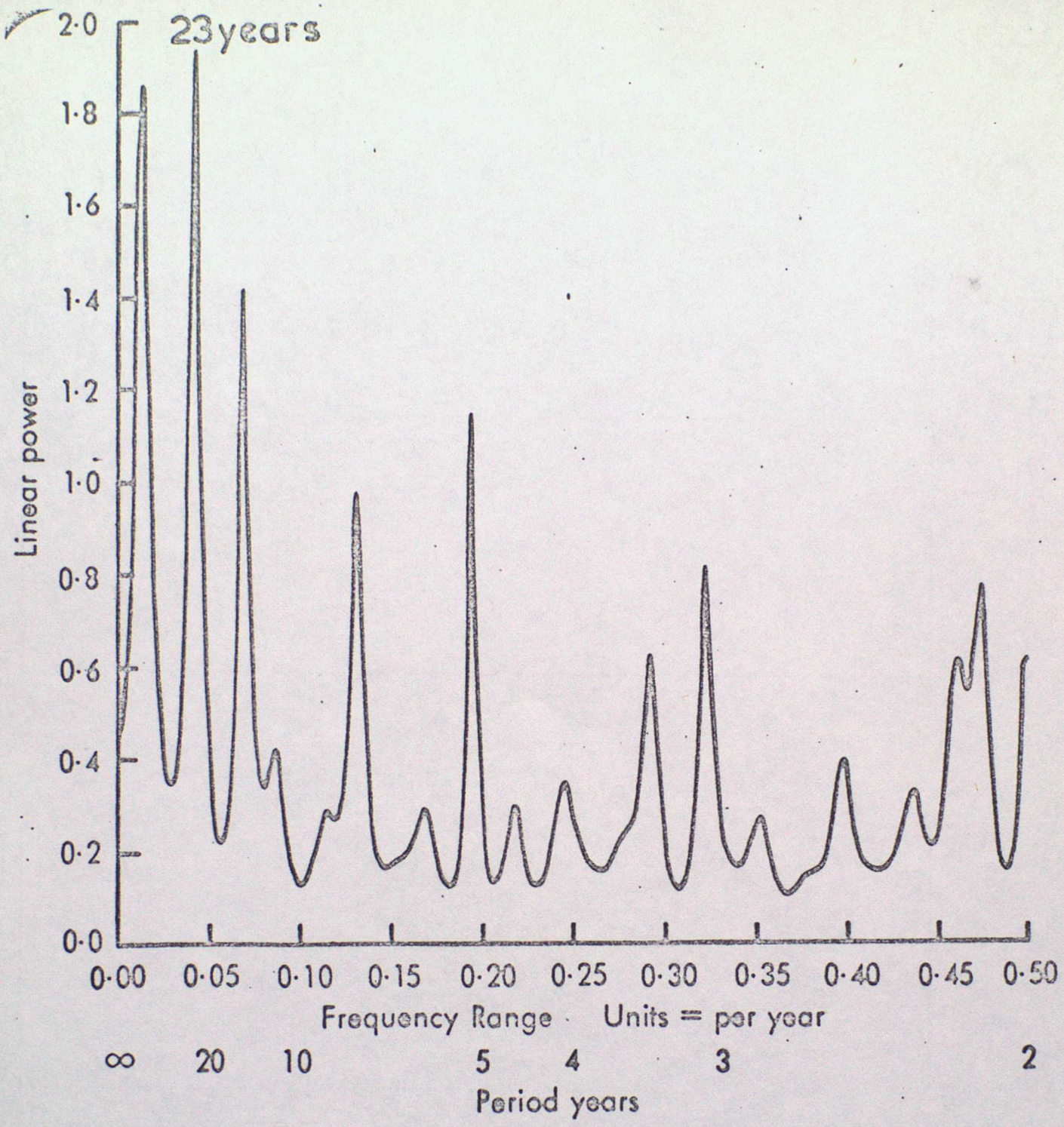
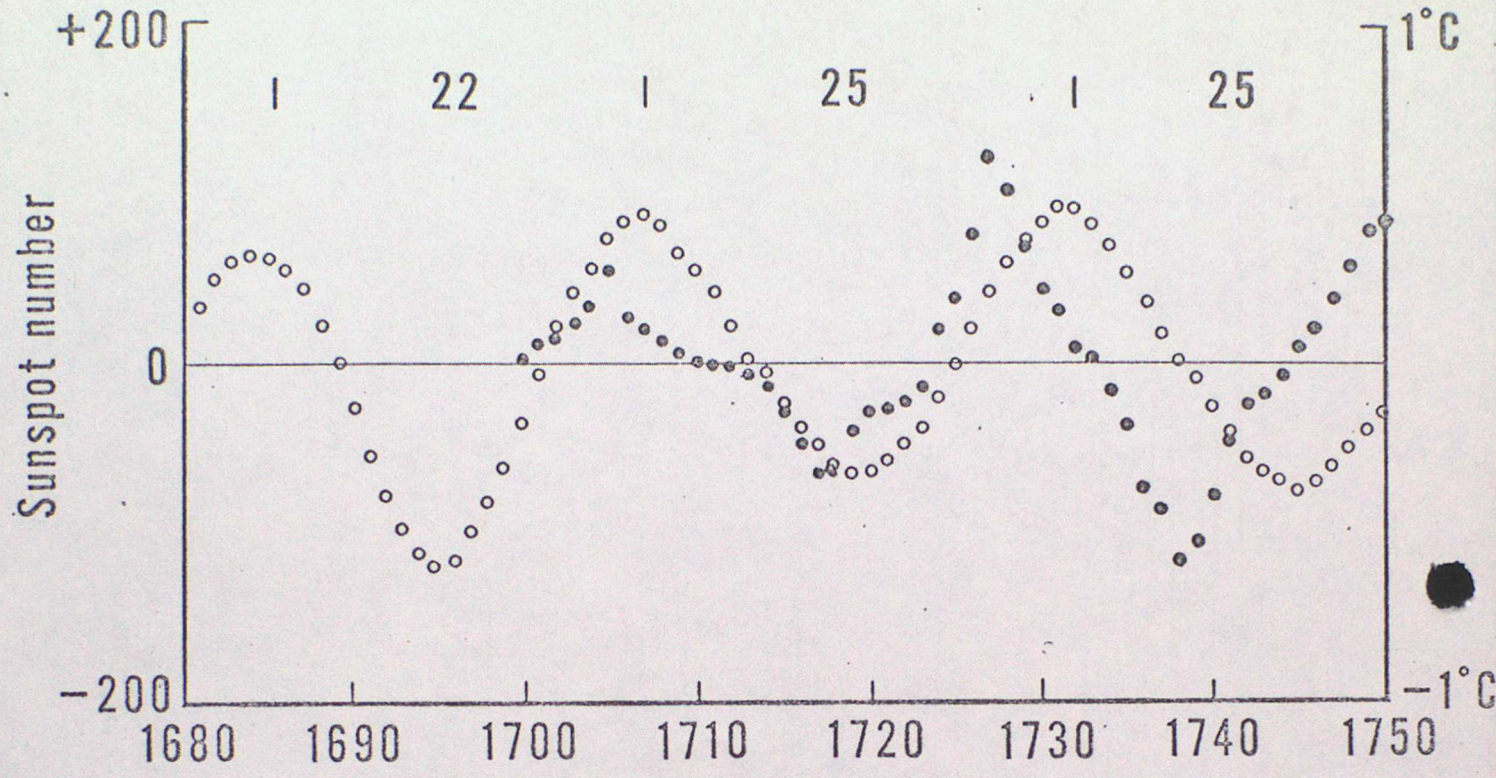


FIG. 17 Power spectrum of annual Central England temperature 1659-1974

FIG. 18A 23 YEAR "CYCLE"



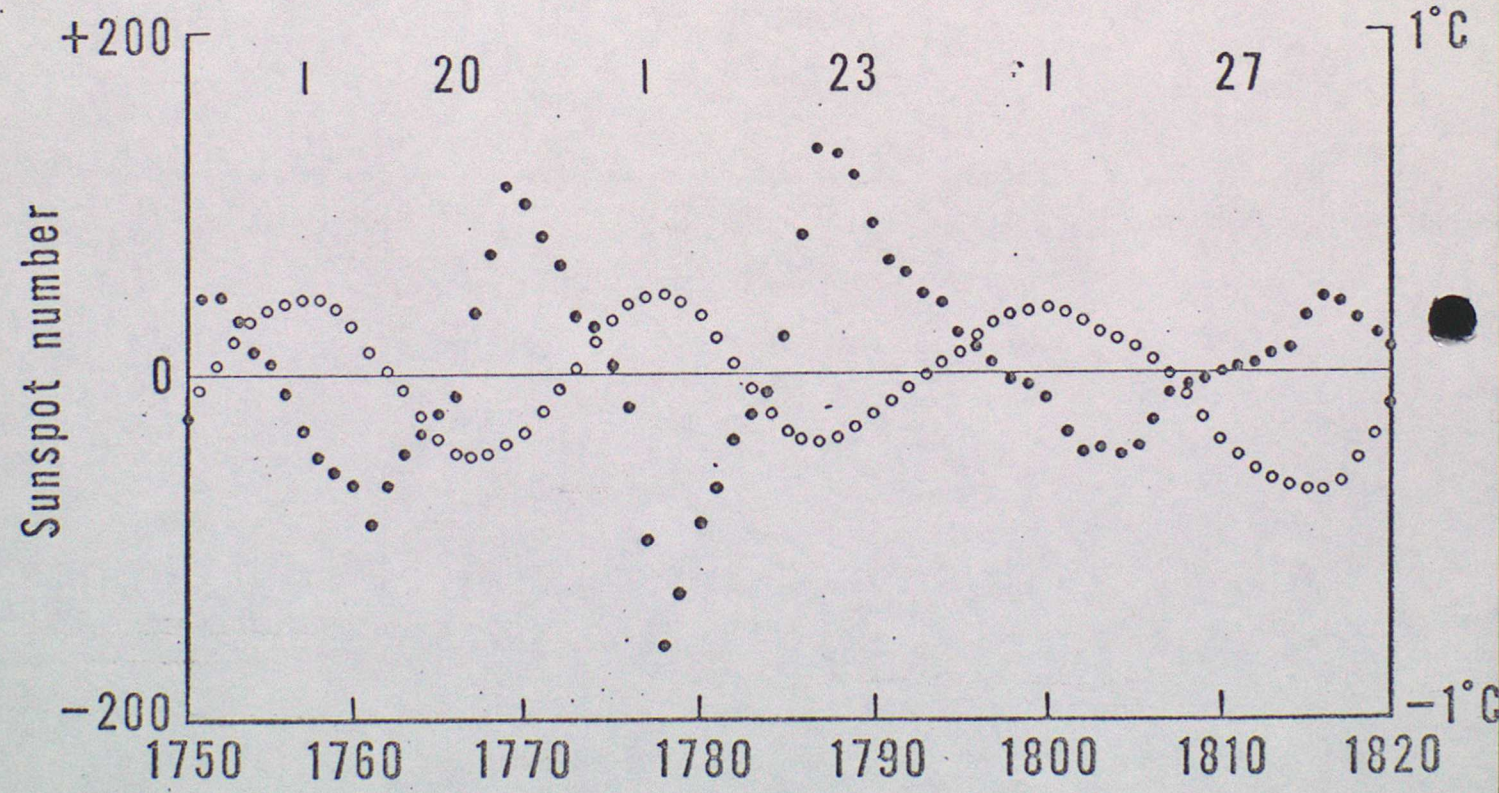


FIG. 18B 23 YEAR "CYCLE" (CONT.)

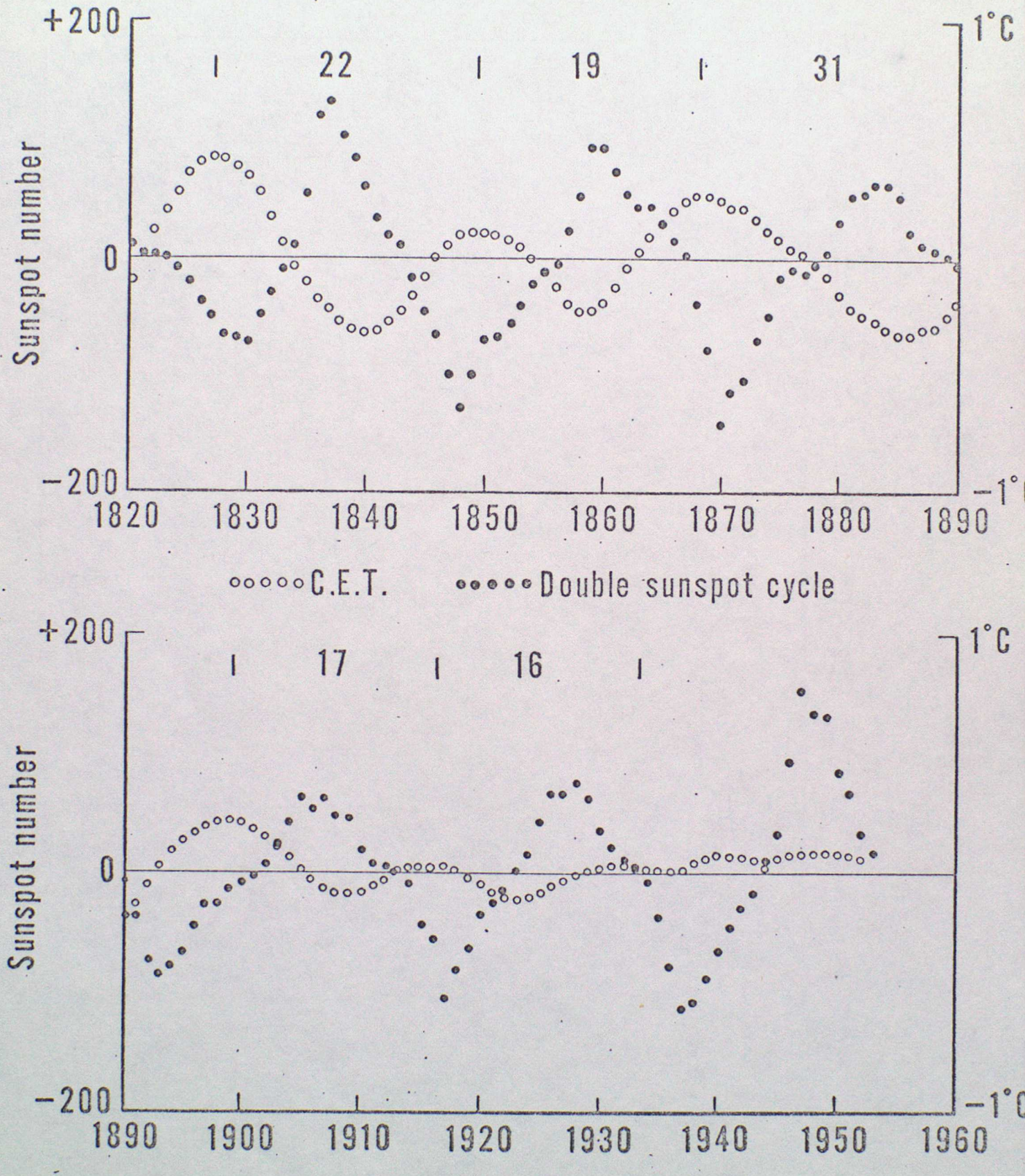


FIG. 19

DAILY C.E.T. POWER SPECTRUM 1911–1915

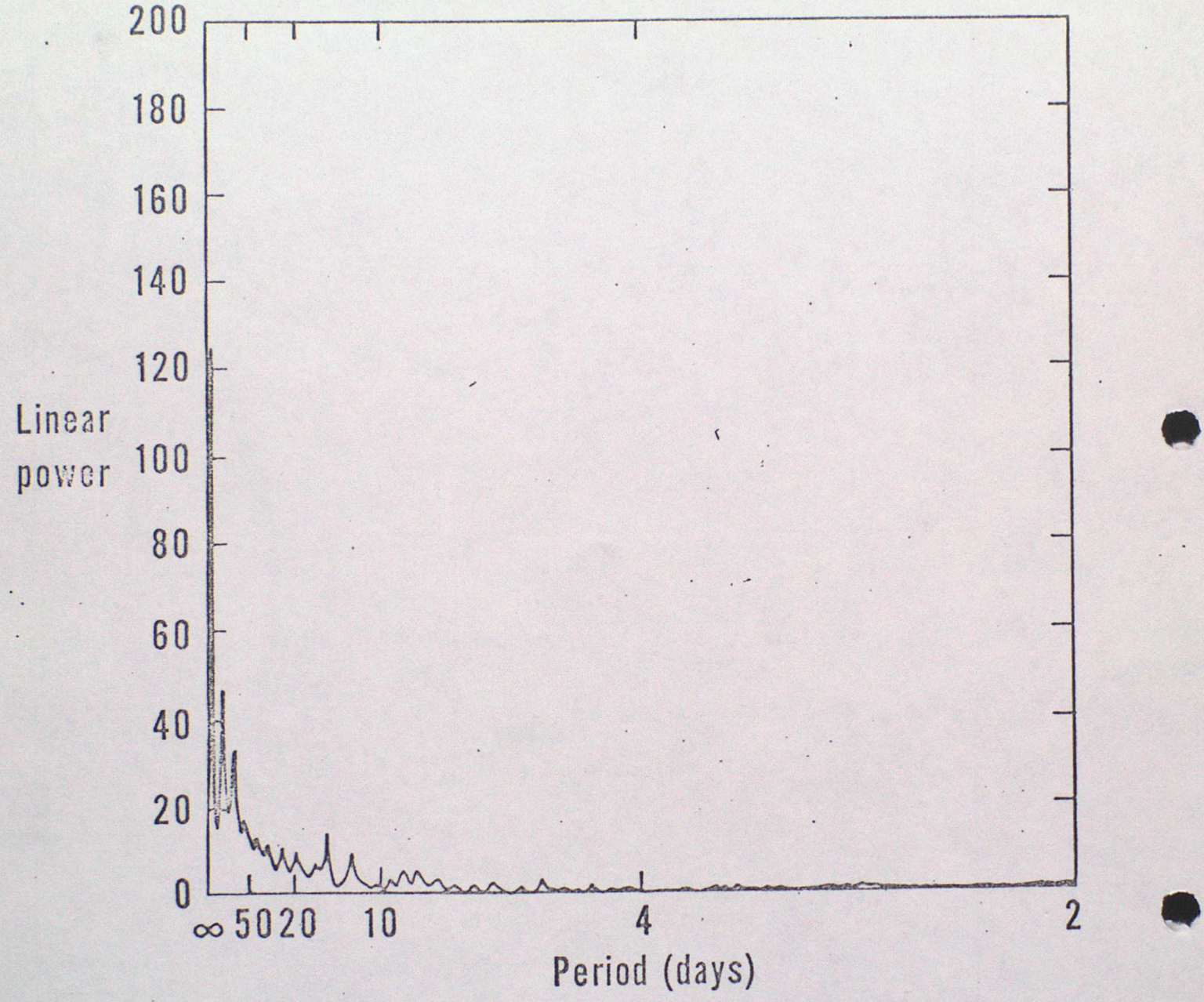


Fig. 20 Ratio of variance on "30 day" to "weather"
A time scales (500 mb ht)

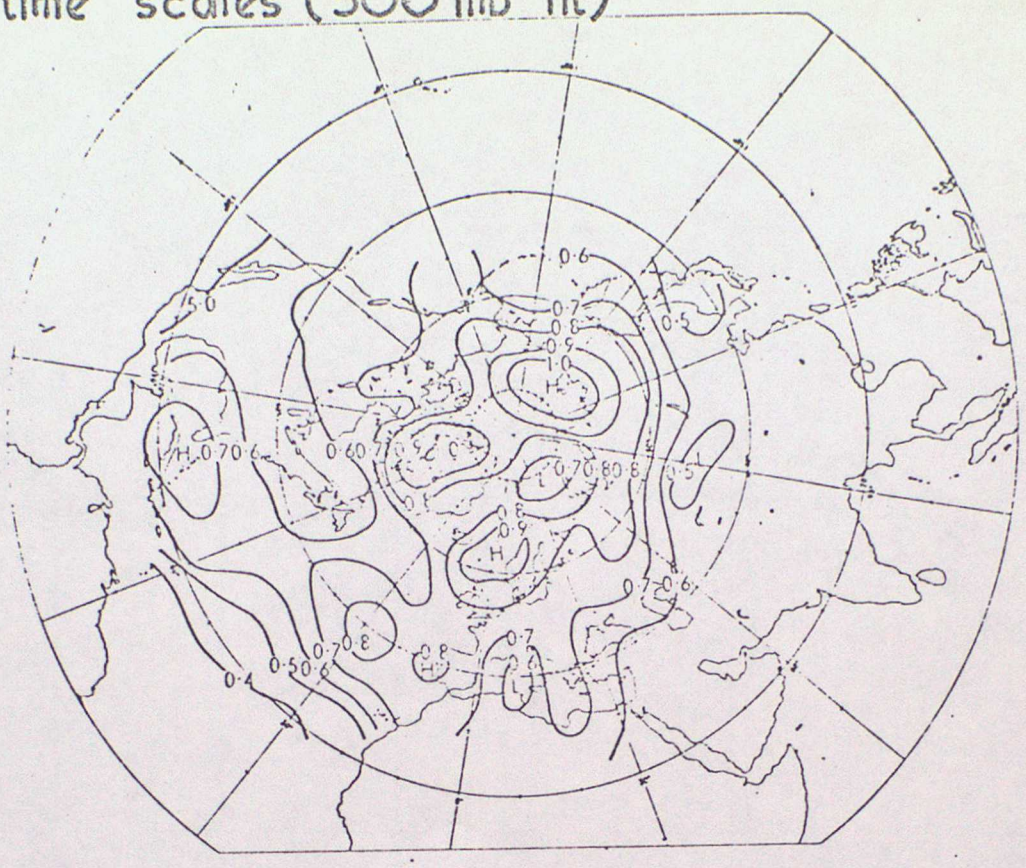


Fig 20 Correlation of 500 mb. height and 500 mb.-
B 1000 mb. thickness

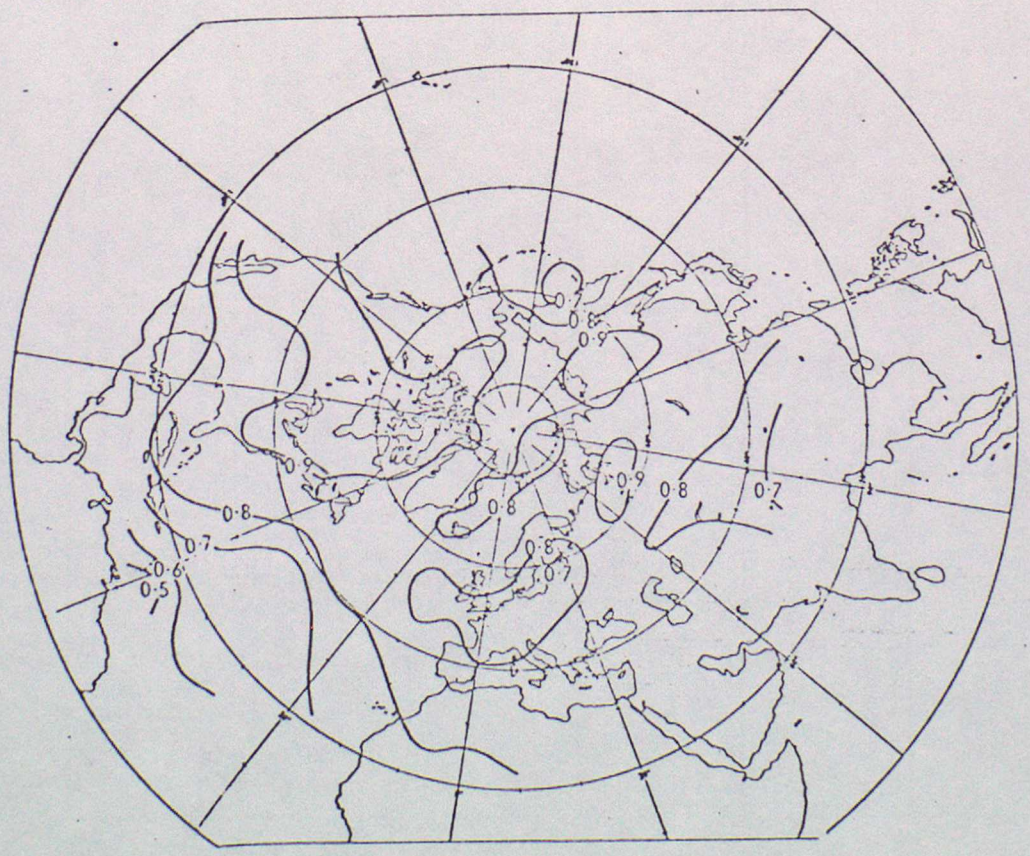
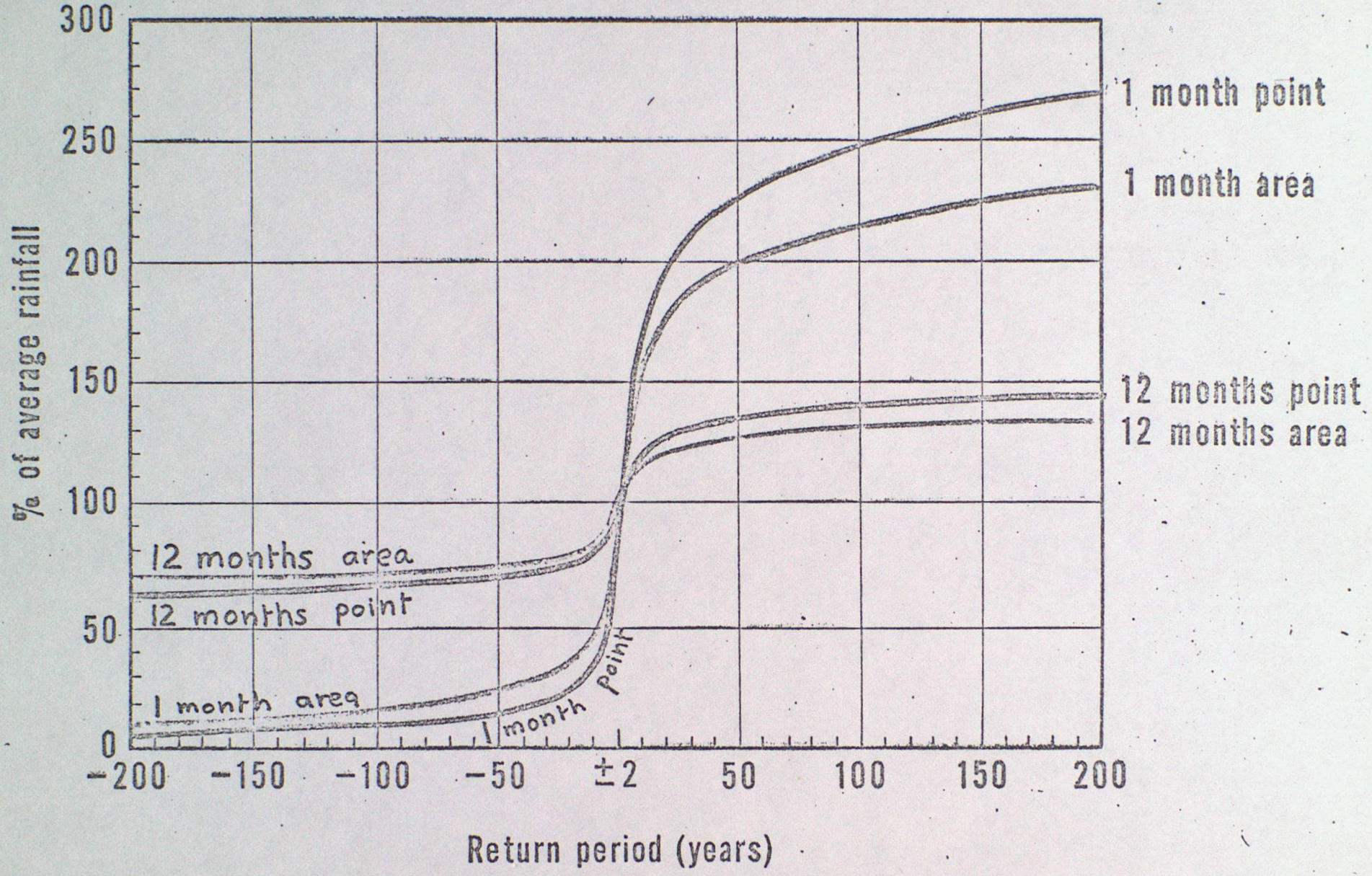


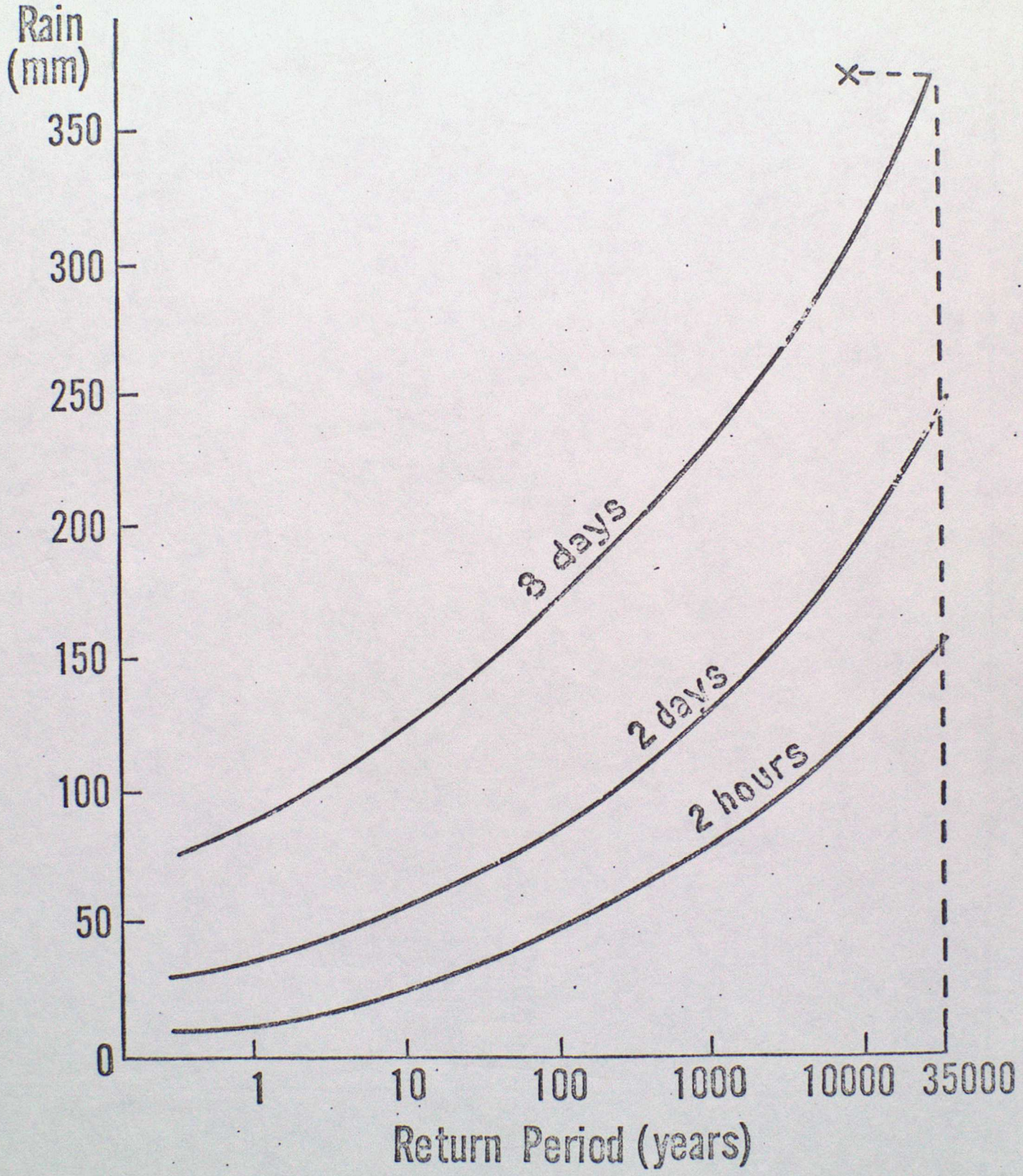
FIG.21 ENGLAND AND WALES RAINFALL



Negative values are rainfall deficiencies

FIG. 22

Typical Rainfall frequency curves for short durations at a point (S. E. England)



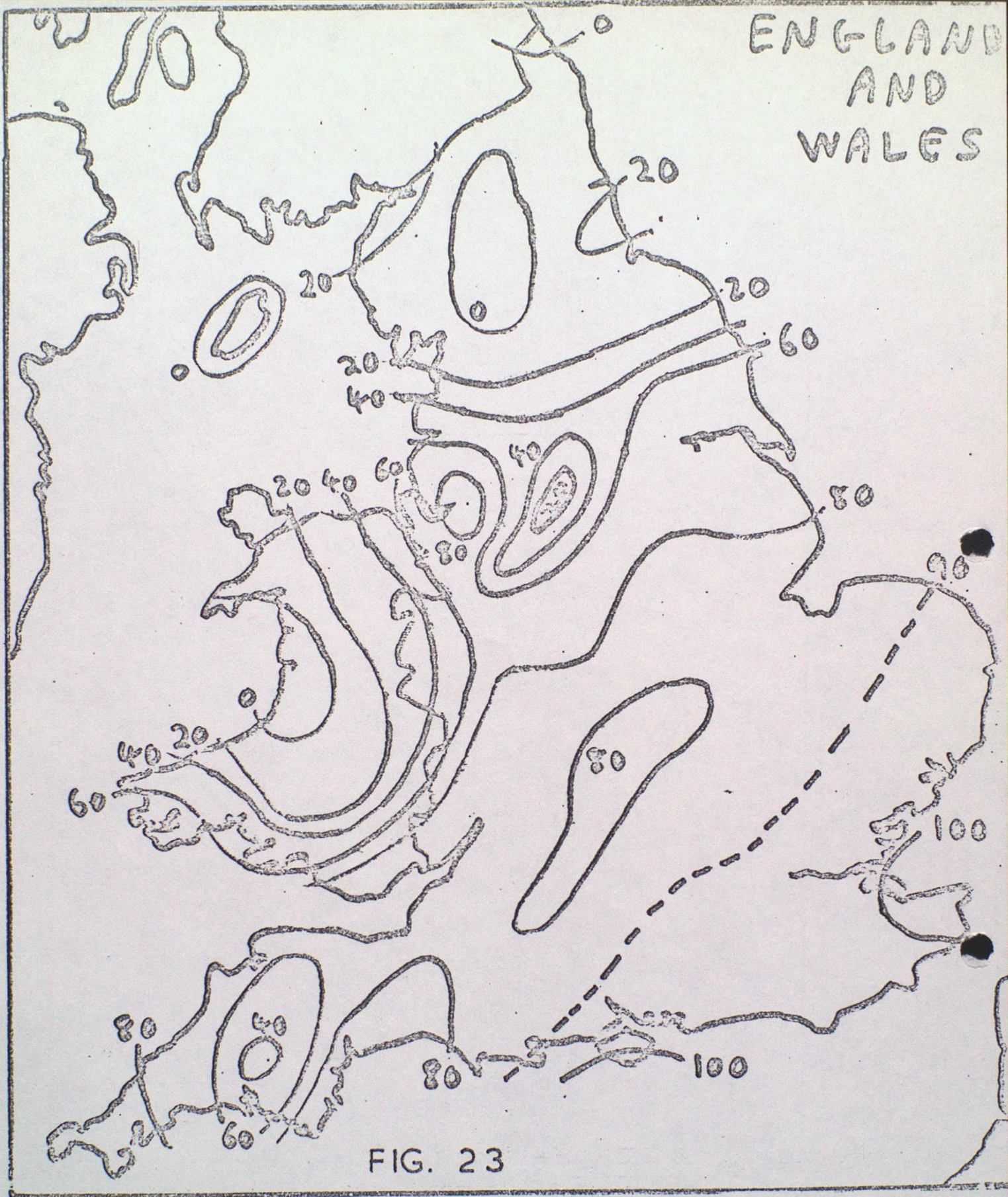


FIG. 23

PERCENTAGE OF YEARS SUITABLE FOR TOWER SILAGE MAKING FROM MAIZE

VS

METEOROLOGICAL OFFICE - ADVANCED LECTURES 1982

THE CLIMATE OF THE WORLD

Lecture 5

Simple Energy-balance Climate Models

by

D J Carson

Met O 13

March 1982

Simple Energy - balance Climate Models

by

D J Carson

1. Introduction

An increasing awareness of man's vulnerability to climate variations and a growing belief that his activities may have a significant influence on the climate have led, over the past decade, to the World Climate Conference in 1979 (WMO,1979) and to the formulation of the World Climate Programme (WMO,1980), one component of which is the programme for research on climate change and variability, the World Climate Research Programme (WCRP). One of the main elements in the WCRP is the development and evaluation of climate models to be used for the study of climate predictability and climate sensitivity.

There is as yet no comprehensive theory of climate to explain its variability, nor are there, at this juncture physical-dynamical models which can simulate adequately the subtleties and complexities of the full climate system. Indeed there exists an almost continuous spectrum of types of so-called climate model, ranging from simple, one-dimensional, equilibrium representations of the vertical radiative processes in the atmosphere to the full complexity of the three-dimensional, time-dependent, global general circulation models. The latter, in which the effects of large-scale disturbances in the atmosphere are included explicitly, are sometimes referred to as comprehensive models in contrast to the simple models in which many more dynamical and physical processes have to be parametrized. There is of course debate about the advantages and deficiencies of each class of model but at present it is probably fair to claim that each type has a rôle to play in furthering our understanding of the real climate system. It should be obvious, for example, that some of the time scales of interest in climate modelling make it both economically and practically impossible to apply the fully comprehensive models in many studies and so the simpler, usually less realistic, models have to be used.

The Meteorological Office has a major commitment to develop its atmospheric general circulation models to produce, ultimately, a comprehensive, coupled atmosphere - land - ocean - cryosphere climate model. The structure and performance of such general circulation models have been discussed, for example, by Dr A Slingo in the complementary lecture series on The General Circulation of the Atmosphere. The aim of this lecture is not to review the state of the art of simple climate modelling (there exists too large a body of literature!) but rather it is to introduce the reader to the concept of such models and to the assumptions made in the simplest examples. Their application to climate-related problems and the somewhat surprising properties of their solutions will also be discussed.

Before describing the models it is worth introducing some terms which are likely to be encountered, increasingly without being explained, in the literature concerned with the design of climate models and their application to sensitivity and predictability studies.

Predictability of Different Kinds

Consider the two questions:

(1) Is climate predictable in some sense or another and at some time range ?

(2) Are man's activities influencing the climate in any way ?

Different modelling methodologies are required to tackle the two, generally distinct, aspects of climate predictability embodied in these questions. Lorenz (1975) recognised these as questions of predictability of the first and second kind, respectively.

Predictability of the First Kind

This is defined as the prediction of the natural variations of the climate system i.e. it is effectively long-range forecasting using the present state of the climate system as an initial condition.

In principle this includes variations from the synoptic scale limit of predictability of a few weeks up to the epochs of glaciations. The nature of climate predictability of the first kind is not understood anywhere in its range. Indeed, a major open question is whether there are any characteristics of the circulation that retain predictability in the range of several weeks to several years. Even if climate predictability of the first kind is possible in principle it remains to be determined how the problem can be approached. The long-range forecasting for a month to a season ahead, which is being researched in the Synoptic Climatology branch (Met O 13), comes into this category at the short-period end of the range. Because such relatively short periods (in climate terms) of prediction are involved at present in such studies, then fully comprehensive general circulation models can be used.

Predictability of the Second Kind

This is defined as the determination of the response of the climate system to changing external influences which might be prescribed or predicted in terms of other dynamical and physical considerations. Such predictions are not concerned directly with the chronological order of occurrence of atmospheric states, but rather with the long-term average statistical properties of the climate (or model).

Most applications of atmospheric general circulation models to climate problems are necessarily predictability studies of the second kind i.e. the typical set of control and perturbation integrations to determine the sensitivity of a simulated climate to prescribed changes in external forcing. A topical example is the study of the impact on the simulated climate of doubling the amount of CO_2 in the atmosphere (see, for example, Mitchell (1981)). It is still an unresolved question whether predictability of the second kind is a meaningful concept or not.

Transitivity, Intransitivity and Almost Intransitivity

An important issue of climate theory is whether the causes of most climate fluctuations are external to the climate system or internal to it, i.e. have past changes been caused, for example, by fluctuations in solar emissions or changes in the earth's orbital parameters (the so-called Milankovich Theory of climate change), or can these changes be explained to a large extent by the intrinsic internal variability within the system, which needs only an external trigger ? This question has been given a more precise mathematical formulation by Lorenz (1968, 1970, 1975) who introduced the concepts of transitivity, intransitivity and almost intransitivity, taken from ergodic theory.

Transitivity. If the very long-term averaged statistics of a climate system (or climate model) are independent of the initial state, then that climate (or model) is said to be transitive.

Intransitivity. If no unique set of such equilibrium climate statistics exists then the climate (or model) is called intransitive, i.e. there are two or more sets of statistical properties and some initial states lead to one set while others lead to another.

Almost Intransitivity. If the climate system (or model) is transitive for statistics taken over an infinite period and yet contains shorter-period self-fluctuations that might be interpreted as climate changes indicative of an intransitive climate system, then it is called almost intransitive. An almost intransitive system could be mistaken for a truly intransitive one if the statistics over long but finite times were identified as those over an infinite period. For example, the existence of glacial and interglacial periods raises the possibility that the climate system is almost intransitive.

There exists the possibility that for fixed external conditions, very different climate states may be implied by finite-interval time averages owing solely to intransitivity or almost intransitivity. The association of a particular climate with a given external state requires an assumption (as yet unfounded) of transitivity of the climate. With any climate model it is very important to determine the sensitivity of its equilibrium state statistics to variations of internal (e.g. initial) conditions. It will be discussed below that some of the simplest climate models are intransitive and do in fact support two steady state solutions corresponding to glacial and interglacial climates.

2. Radiative-convective Thermal Equilibrium Models (Vertical Column Energy-balance Models)

2.1 Introduction

This class of one-dimensional models computes the annually and horizontally averaged, global, surface and atmospheric temperatures. The temperature profile is determined essentially from a balance between the incoming short-wave (or solar) and the outgoing long-wave (alternatively called terrestrial, infrared or thermal) radiative fluxes, which, in turn, depend upon specified surface properties and vertical distributions and optical properties of clouds, aerosols and of the most important of the optically active gases such as water vapour, carbon dioxide and ozone. A method of convective adjustment is used to ensure that realistic vertical temperature profiles are obtained.

The first radiative-convective models for the earth's atmosphere were developed by Manabe and Strickler (1964) and Manabe and Wetherald (1967). Further variations of the approach have also been developed and are discussed in the comprehensive review by Ramanathan and Coakley (1978). Radiative-convective modelling is a technique commonly employed in studies of stellar and planetary atmospheres.

The paper by Manabe and Wetherald (1967) is to some extent a repeat of the computation of radiative-convective equilibrium of the atmosphere which was discussed in the earlier paper by Manabe and Strickler (1964). The main difference is that Manabe and Wetherald specify a given distribution of relative humidity instead of a given distribution of absolute humidity as used in Manabe and Strickler. The former assumption is thought to be the more realistic. The prime objective in both cases was to formulate and test a method to incorporate radiative transfer into a general circulation model.

Note that with respect to the formulation of the radiative transfer processes in the single column the radiative-convective models are not necessarily simple. Because so much else is parametrized in such a simple way the combination of high vertical resolution and the small demand made on computing resources enables the radiative processes to be modelled in a very detailed fashion. It is not the purpose of this lecture to examine the subtleties and complexities of the actual schemes used by Manabe and his colleagues; for this information the reader is referred to the original papers.

Further objectives of such modelling were:

(1) To determine the influence of various factors such as the solar constant, cloudiness, surface albedo and the various atmospheric absorbers in maintaining the existing thermal structure of the atmosphere.

(2) To determine the rôles of the above factors in maintaining the observed climate at the earth's surface.

2.2 Model assumptions and structure

a. Pure radiative equilibrium

In order to prepare for the study of thermal equilibrium with convective adjustment, calculations were first done for the simple case of purely radiative equilibrium, in which the equilibrium state was approached as an asymptotic steady state of an initial value problem. Equilibrium is considered to have been reached when the temperature change at each level over the time-step used in the model is less than a specified small value.

The following constraints are also satisfied in the equilibrium state:

(1) At the earth's surface, the net upward flux of long-wave radiation is equal to the net downward flux of solar radiation. This guarantees radiative equilibrium at the surface.

(2) At the top of the atmosphere the net upward flux of long-wave radiation is equal to the net downward solar radiation. This guarantees the radiative equilibrium of both the atmosphere and the earth as a whole.

The unrealistic temperature profiles which result when radiation alone is acting have been described by Dr A Slingo in the complementary lecture series on the General Circulation of the Atmosphere and so there is no need to discuss them further here.

b. Radiative-convective equilibrium

The procedure of convective adjustment is to modify the lapse rate to the critical lapse rate whenever that critical lapse rate is exceeded in the course of the integration. The critical lapse rate generally adopted is 6.5 K km^{-1} which was assumed to be a reasonably realistic value for the observed tropospheric lapse rate of temperature. The process of convective adjustment will transfer heat energy from the surface into the atmosphere and thereby permit more realistic temperatures to occur throughout the troposphere.

The requirements for radiative-convective equilibrium are:

(1) At the top of the atmosphere, the net incoming solar radiation should be equal to the net outgoing long-wave radiation.

(2) No temperature discontinuity should exist.

(3) Free and forced convection, and mixing by large-scale eddies, prevent the lapse rate exceeding a critical value of 6.5 K km^{-1}.

(4) Whenever the lapse rate is subcritical, the condition of local radiative equilibrium is satisfied.

(5) The heat capacity of the earth's surface is zero.

(6) The atmosphere maintains the given distribution of relative humidity. (This is a requirement introduced in Manabe and Wetherald (1967)).

Figure 1 gives a simple schematic representation of the iterative procedure used to obtain the state of radiative-convective equilibrium. Both 18 and 9 atmospheric levels were used in the computations such that the pressure thickness of the layer is thin both near the surface and near the top of the atmosphere.

Mathematical Note: r (the humidity mixing ratio) $= \dfrac{0.622\, e}{p - e}$

where e is the water vapour pressure, p is the atmospheric pressure.

$$h \text{ (the relative humidity)} = \frac{e}{e_s}$$

where e_s is the saturation water vapour pressure.

Therefore, r can be written

$$r = \frac{0.622\, h\, e_s(T)}{p - h\, e_s(T)}$$

where e_s as a function of temperature, T, can be evaluated from the Clausius-Clapeyron relationship,

$$\frac{de_s}{dT} = \frac{0.622\, L\, e_s}{R_d T^2}$$

where L is the latent heat of evaporation and Rd is the specific gas constant for dry air.

c. Details of the radiation scheme

The reader is referred to Manabe and Strickler (1964) for details of:

(1) The computation of the flux of long-wave radiation.

(2) The computation of the depletion of solar radiation.

(3) The determination of mean absorptivity and emissivity.

d. Assumptions made about standard distributions of atmospheric absorbers

The following constituents and their vertical distributions are adopted unless otherwise specified.

Water vapour — The relative humidity is assumed to be a linear function of pressure throughout most of the atmosphere, as illustrated in Figure 2. The line was chosen to fit the data quoted in the Figure legend and a constraint of a minimum value of the humidity mixing ratio of 3×10^{-6} is enforced.

Carbon dioxide — The mixing ratio of carbon dioxide in the atmosphere is assumed constant at 0.0456 % by weight (300 ppm by volume).

Ozone — The adopted vertical distribution of ozone is shown in Figure 3.

The verticaldistributions of gaseous absorbers are those appropriate for 35°N in April. Insolation is assumed to be the annual mean value for the hemisphere.

Clouds — The characteristics of the clouds implied in the model are listed in Table 1. The albedo of the earth's surface is assumed to be 0.102.

Cloud	Height (km)	Amount	Albedo
High	10.0	0.228	0.20
Middle	4.1	0.090	0.48
Low: Top	2.7	0.313	0.69
Bottom	1.7		

Table 1. Cloud characteristics used in the radiative-convective equilibrium model of Manabe and Wetherald (1967).

2.3 Some results from radiative-convective equilibrium models

Figure 4, from Manabe and Wetherald, shows the purely radiative equilibrium profile obtained with the assumption of fixed absolute humidity (as used by Manabe and Strickler) and that obtained with the assumption of fixed relative humidity. Note that with fixed relative humidity there is a very sharp decrease with height near the surface and that the temperature of the troposphere is generally much lower than that obtained with fixed absolute humidity. The radiative-convective equilibrium profile for the case with fixed relative humidity is also shown and demonstrates quite clearly the role of convective adjustment in establishing and maintaining the existing distribution of atmospheric temperature (Cf. Figure 6 of Slingo (1982)).

Figure 5, from Manabe and Strickler, shows the approach to pure radiative equilibrium and radiative-convective equilibrium from both initially warm and initially cold isothermal states, respectively. This figure indicates that the final states may be regarded as steady states.

A wide range of experiments has been done to examine the sensitivity of the radiative-convective equilibrium temperature profiles obtained, for example, with different values of the solar constant and the surface albedo and different combinations and vertical distributions of the main atmospheric absorbers such as water vapour, CO_2, O_3 and clouds. Of particular, topical interest are the experiments of Manabe and Wetherald which examined the response of the surface temperature to changes in the atmospheric CO_2 content.

Figure 6, from Manabe and Wetherald, shows the vertical distributions of equilibrium temperature corresponding to three different CO_2 contents, viz. 150, 300 and 600 ppm. Note:

(1) The larger the mixing ratio of CO_2 the warmer is the equilibrium temperature of the earth's surface and troposphere. This results primarily from an enhanced absorption of the surface long-wave radiation by the tropospheric CO_2 but is also partly due to the effective increase in downward radiation emitted by stratospheric CO_2.

(2) The larger the mixing ratio of CO_2, the colder is the equilibrium temperature of the stratosphere. This results from the increase in long-wave radiation emitted to space by CO_2 in the stratosphere.

(3) Relatively speaking, the dependence of the equilibrium temperature of the stratosphere on CO_2 content is much larger than that of the tropospheric temperature.

Tables 2 and 3 show the equilibrium temperature of the earth's surface corresponding to different CO_2 contents and also the change of the equilibrium surface temperature corresponding to the specified changes in CO_2. Values for both fixed absolute humidity and fixed relative humidity cases are given. Note that the equilibrium temperature of the latter is almost twice as sensitive to the change in CO_2 as that of the former due to the adjustment of water vapour content to the temperature variation of the atmosphere. The changes resulting from doubling the CO_2 content are in reasonable agreement with those obtained in subsequent experiments with general circulation models. Table 4, from Mitchell (1981), gives a comparison of estimates of the change in global mean surface temperature obtained for specified changes in the CO_2 content from single-column, radiative-convective equilibrium models (including the results of Manabe and Wetherald (1967)) with the changes obtained from corresponding general circulation model integrations.

	Average cloudiness		Clear	
CO_2 content (ppm)	Fixed absolute humidity	Fixed relative humidity	Fixed absolute humidity	Fixed relative humidity
150	289.80	286.11	298.75	304.40
300	291.05	288.39	300.05	307.20
600	292.38	290.75	301.41	310.12

Table 2. Equilibrium temperature of the earth's surface (K) for different values of the CO_2 content of the atmosphere, from Manabe and Wetherald (1967).

Change of CO_2 content (ppm)	Fixed absolute humidity		Fixed relative humidity	
	Average cloudiness	Clear	Average cloudiness	Clear
300 ⟶ 150	-1.25	-1.30	-2.28	-2.80
300 ⟶ 600	+1.33	+1.36	+2.36	+2.92

Table 3. Change of equilibrium temperature of the earth's surface corresponding to various changes of CO_2 content of the atmosphere, from Manabe and Wetherald (1967).

2.4 Limitations of the radiative-convective equilibrium models

(1) The most important limitation is that results from these models are mostly of academic interest, since they do not give any information about regional and latitudinal temperature changes.

(2) Many of the model parameters (cloud amounts, surface albedo, relative humidity, critical lapse rates, etc) are prescribed on the basis of present day conditions which may not apply for large departures from present day conditions. Radiative-convective models cannot be applied for large perturbations from present conditions.

(3) The model results depend very critically on the details of the parametrizations. Also like many other models, particularly the simpler models, these models are probably too deterministic.

It is of course very important that the limitations of all the models, especially the simpler, cruder models, be borne in mind when their results are being interpreted.

3. Horizontally Varying Energy-balance Models

3.1 Introduction

In contrast to the radiative-convective equilibrium models of the previous section, this class of one-dimensional, energy-balance models treats the vertical structure with much less detail than the horizontal structure. There exist many types of model of varying degrees of complexity within this class but here we shall concentrate on those which presume to compute the zonally-averaged surface temperature as a function of latitude only, obtained effectively as a balance between incoming solar and outgoing terrestrial radiation. Budyko (1969) and Sellers (1969) are generally acknowledged as having proposed independently the first such models and it is now common practice to refer to the class of derivatives and extensions of their approaches as Budyko - Sellers type models.

Table 4. Estimates of global mean changes in temperature and precipitation due to increasing atmospheric carbon dioxide.

Source	3-dimensional model							
	Topography	Solar Radiation	Cloud	Ocean	CO_2	PPTN	T*	1-D model T*
Manabe and Wetherald, 1975	Idealized	Annual mean	Fixed	Swamp	2	7%	2.93	2.36 †
Manabe and Wetherald, 1980	Idealized	Annual mean	Model Dependent	Swamp	2	7%	3.0	2.0
					4	12%	5.9	4.1
Manabe and Stouffer, 1980	Realistic	Seasonal	Fixed	Fixed depth	4	6.7%	4.1	
Present work								
(a) 2 x CO_2 + 2K	Realistic	Seasonal	Fixed	Present day + 2K	2	5%	2.25	
(b) Estimate assuming no change in global mean net surface heat balance				Present day + 1.65K	2	3.5%	1.9	1.7

† Manabe and Wetherald (1967)

This table is taken from Mitchell (1981).

There is incontrovertible evidence of the existence of large fluctuations in past climate. There is also debate that the impact of man's activities may possibly initiate a significant response in the climate system. Therefore there is concern about and interest in the question of the stability of the global climate to small perturbations of any sort. This led Budyko (1969) and Sellers (1969) to develop their simple models of the earth's climate, based on the mean annual balance of heat entering and leaving latitudinal strips of the earth-atmosphere system, for the purpose of testing the sensitivity of the equilibrium state of the climate to changes in the solar radiative heating.

Schematically, the heat balance for the i-th strip in equilibrium may be written:

$$\left\{\begin{array}{c}\text{Absorbed solar} \\ \text{radiation}\end{array}\right\}_i - \left\{\begin{array}{c}\text{Outgoing infrared} \\ \text{radiation}\end{array}\right\}_i = \left\{\begin{array}{c}\text{Horizontal divergence of} \\ \text{thermal energy flux}\end{array}\right\}_i$$

Each of the above terms is parametrized by some theoretical or semi-empirical formula, usually in terms of the sea-level temperature field. All of the equations are then solved simultaneously to give the surface temperature field. Note that the balance equation assumes that there is no heat or moisture storage in the oceans, land or atmosphere and hence no long-term climatic trend. There is also the gross assumption that the character of the climate of the whole column of the earth-atmosphere system is embodied in a single variable, the surface temperature. Although the various particular studies differ in the exact details of their parametrizations of the above components of the balance it is generally believed that they are at least qualitatively equivalent.

Although these models are highly simplified their solutions to particular problems have a rather complicated structure. Indeed it will be shown that a multiplicity of solutions is obtained for the present value of the solar constant. The interesting and, arguably, unexpected character of such solutions, their mathematical stability properties and the consequent implications for our appreciation of climate are the main reasons why an increasingly large amount of study has been devoted to this class of simple climate model since the appearance of the papers by Budyko (1969) and Sellers (1969).

3.2 The model of Budyko (1969)

In the first instance let us consider the model and results of Budyko (1969). The simplified, integrated thermodynamic energy equation for the whole earth-atmosphere system is represented in terms of annually and zonally averaged variables and equations, thus

$$R + \operatorname{div} \underset{\sim}{F} = 0 \qquad (1)$$

where R is the net radiation for a column of the earth-atmosphere system at a given latitude and $\operatorname{div} \underset{\sim}{F}$ represents the net gain or loss of heat, i.e. it represents the divergence of the net energy flux, as a result of the atmospheric and hydrospheric circulations for that latitude

The net radiation is expressed as

$$R = Q(1-\alpha) - I \qquad (2)$$

where Q is the annual mean solar radiation at the outer boundary of the atmosphere, α is the planetary albedo (or reflectivity) and I is the net terrestrial radiation to space, all for a given latitude.

An analysis of radiation and temperature data led Budyko to propose and use the empirical formula

$$I = a + \beta T_o - (a_1 + \beta_1 T_o) n \qquad (3)$$

where I is the outgoing radiation in kcal cm^{-2} $month^{-1}$,
T_o is the temperature at the earth's surface in ^oC,
n is the cloudiness in fractions of unit, and
a, β, a_1, β_1 are dimensional coefficients with values
$a = 14.0$, $\beta = 0.14$, $a_1 = 3.0$ and $\beta_1 = 0.10$.

For the earth as a whole (i.e. with div $\underset{\sim}{F} \equiv 0$) Budyko found that with $n = 0.50$ and constant $\alpha = 0.33$, a change of solar radiation by 1% results in a temperature change of 1.5K. He cites a corresponding estimate from Manabe and Wetherald (1967) as giving a 1.2K change in surface temperature as a result of a solar radiation change of 1%. Budyko argues that the long time-scales involved in any climate response resulting from radiation variations require that changes in the earth's albedo should be taken into account which are due to the expansion or reduction of the ice-cover at the surface. The parametric coupling that results between α and T_o provides a strong positive feedback between ice-cover and temperature and we shall see later that it is this empiricism which is responsible for many of the surprising results.

The flux divergence term in Equation (1) is also specified empirically. Budyko derived a relationship which enabled him to represent this term quite simply in terms of the difference between the local zonal mean and the planetary mean temperature, thus

$$\text{div } \underset{\sim}{F} = -\beta (T_o - T_p) \qquad (4)$$

where T_p is the globally averaged, annual mean surface temperature and β is an empirically derived coefficient equal to 0.235 kcal cm^{-2} $month^{-1}$ K^{-1}.

From (1) - (4) we obtain the following equations:

$$T_o = \frac{Q(1-\alpha) - a + a_1 n + \beta T_p}{\beta + \beta_1 - \beta_1 n} \qquad (5)$$

$$T_p = \frac{Q_p(1-\alpha_p) - a + a_1 n}{\beta - \beta_1 n} \qquad (6)$$

where Q_p, α_p are the corresponding planetary values of incident radiation and albedo, respectively. Budyko used these to derive the average latitudinal annual mean temperatures corresponding to present climatic conditions for the Northern Hemisphere. Q and Q_p were computed corresponding to a solar constant of 1.92 cal cm^{-2} min^{-1}. In the calculation, the influence on the temperatures of deviations of cloudiness values from the mean planetary value of 0.50 is neglected.

Based on observational data Budyko chose the following specification of albedo:

$$\alpha = \begin{cases} 0.32 & 0 \leq \varphi \leq 60^\circ \quad (\varphi \text{ is latitude}) \\ 0.50 & \varphi = 70^\circ \\ 0.62 & \varphi = 80^\circ \end{cases}$$

Budyko's calculation of contemporary average latitudinal distribution of temperature is shown in Figure 7. The computed values of $\overline{T}_o$ are in good agreement with the observed values, $\overline{T}$, and this encouraged Budyko to use his model to study the effects of variations in the radiation on the Earth's thermal regime and glaciations.

Budyko asserts that the southern boundary of the Arctic ice corresponds to a mean latitude of 72°N and adopts an ice-albedo-temperature coupling such that the ice-cap expands or contracts to coincide with the surface whose temperature is less than or equal to the temperature observed now at 72°N i.e. about -10°C. The -10°C isotherm therefore determines the extent of the ice-covered area which is given an albedo of 0.62 with a discontinuity and drop to 0.50 at its southern boundary. Budyko then computed the zonal mean temperatures and hence the position of the edge of the ice-sheet for effectively different values of the solar constant. The results are shown in Figure 8 where the lines $\overline{T}_{1.0}$ and $\overline{T}_{1.5}$ correspond to the temperature distributions for a 1% and a 1.5% decrease in the incoming solar radiation. It is assumed that the relative decrease in radiation is the same at all latitudes. Figure 9 shows the values obtained for the mean planetary temperature, $\overline{T}_p$ and the mean latitude to which glaciation extends, φ_o, as a function of the relative radiation changes.

The most striking feature of Figure 9 is the implication that only a small per cent (about 1.6%) drop in the solar constant is required for the earth to ice over completely. When radiation decreases by 1.6% the ice-cover reaches the mean latitude of about 50°N after which it starts shifting towards lower latitudes up to the equator as a result of self-development. Simultaneously, $\overline{T}_p$ drops sharply to values several tens of degrees below zero.

3.3 The model of Sellers (1969)

Sellers' starting point is also the integrated energy-balance equation for the earth-atmosphere system for a given latitude belt. His dependent variable is the average annual sea-level temperature, $\overline{T}_s$, in each 10° latitude belt. Again, as in Budyko (1969), several empirical and relatively crude relationships are involved in the model. Using the same definitions and notation as was used above for Budyko's model, the components of Sellers' model can be expressed as follows:

Net terrestrial radiation to space:

$$I = \sigma \overline{T}_s^4 \left[1 - m \tanh \left(19 \overline{T}_s^6 \times 10^{-16} \right) \right] \qquad (7)$$

where σ is the Stefan-Boltzmann constant $(1.356 \times 10^{-12}$ ly $s^{-1} K^{-4})$, and m is an atmospheric attenuation coeffient, taken to be 0.5 for present conditions. Equation (7) is based on an empirical fit to data.

Albedo:

$$\alpha = \begin{cases} b - 0.009 \overline{T}_o & \overline{T}_o < 283.16 K \qquad (8a) \\ \alpha_d = b - 2.548 & \overline{T}_o \geq 283.16 K \qquad (8b) \end{cases}$$

with the constraint that $0.25 \leq \alpha \leq 0.85$ for all T_o,

where $T_o = T_s - 0.0065Z$ is the average surface temperature, Z is the average surface elevation in metres, and b is an empirical coefficient. Values of Z and b are tabulated by Sellers for each latitude belt of 10° latitudinal span.

As in Budyko's model, it is assumed that variations in α are associated mainly with variations in surface snow cover and Equation (8a) is valid only as long as $T_o < 10^oC$ and $\alpha < 0.85$. The planetary albedo α_d of (8b) is used for higher temperatures i.e. in the absence of snow or ice cover. Equation (8a) implies an increase in α by 0.009 for every 1K drop in T_o. The coefficient b was selected so that the equation would fit observed data. The effects of variations in cloud cover on albedo were ignored.

The flux-divergence term:

Sellers identifies three separate components making up the total net flux of heat out of the belt at a given latitude. They are due to, respectively, the net fluxes of water vapour and sensible heat by the atmospheric circulation and sensible heat by ocean currents.

Let F_o be the meridional sensible heat transport due to ocean currents,
F_A the corresponding sensible heat transport due to atmospheric motion,
and F_q the meridional transport of latent heat energy (q is specific humidity),

then Sellers adopts a parametrization which relates these fluxes to the horizontal mixing by large-scale eddies and to a vertically integrated mean meridional wind $\langle v \rangle$ which, in turn, is parametrized as an empirically determined function of the North-South temperature gradient, thus

$$F_o \propto -K_o \frac{\partial T_s}{\partial y} \tag{9a}$$

$$F_A \propto -K_A \frac{\partial T_s}{\partial y} + \langle v \rangle T_s \tag{9b}$$

$$F_q \propto -K_q \frac{\partial q(T_s)}{\partial y} + \langle v \rangle q(T_s) \tag{9c}$$

where K_o, K_A and K_q are effective eddy diffusion coefficients.

An implicit important assumption here is that average atmospheric temperatures and specific humidities and ocean temperatures are proportional to their respective sea-level values. Further the actual specific humidity is replaced by the saturation specific humidity which enables it to be related to the temperature through the Clausius-Clapeyron relationship (see, for reference, the Mathematical Note of Section 2.2b). The poleward transport of water vapour and sensible heat (equations (9b) and (9c)) are assumed to consist of two parts, one associated with a mean meridional motion and the other with large-scale eddies or cyclones and anticyclones. The inclusion of the mean meridional motion is necessary in order to avoid having to deal with negative diffusivities. Values of the K's are tabulated by Sellers at 10° intervals for each latitude circle.

Sellers' system of equations leads to a second-order equation, effectively in T_o, with only one of the roots giving realistic temperatures. The model was first used to reconstruct the present state of the earth-atmosphere system. The reader is referred to Sellers (1969) to verify that the model specifies present conditions very well. Sellers then applied his model to several different problems, two of which are considered here.

a. Modification of the polar ice caps

A number of different combinations fo albedo were used, with α at one or both poles (70 - 90°) being set equal to and held fixed at either 0.50 or 0.40 (with the exception of 70 - 80°S where a minimum value of 0.44 was imposed). The results, shown in Figure 10, have a number of interesting features, viz,

(1) Cases 1 and 2. Removal of north polar ice results in an increase in temperature poleward of 70°N by at most about 7K. Temperature would rise by about 1K in the tropics and 1-3K near the South Pole.

(2) Cases 3 and 4. Removal of antarctic ice produces a temperature increase there of 12-15K with a rise in the Arctic of about 4K.

(3) Cases 5 and 6. Removal of both polar ice-caps yields temperature increases of 7-10K in the Arctic and 13-17K in the Antarctic. Tropical warming does not exceed 2K.

Therefore albedo changes at either pole have global repercussions. An interesting result, comparing the curves for Cases 1 and 3, is that reducing the albedo south of 70°S to 0.50 would, according to this model, produce a larger temperature rise in the Arctic than would a reduction of the albedo north of 70°N to 0.50. The greater effects of changes to the albedo in the Antarctic than in the Arctic are simply the result of the Antarctic's higher initial albedo. Others have argued that the extent of sea-ice in the Southern Hemisphere has a much more significant effect on the global circulation than the extent of sea-ice in the Northern Hemisphere.

b. Variations in the solar constant

Sellers examined the sensitivity of his models to variations in the solar constant and his calculations corroborate, qualitatively at least, the implications from Budyko's model that a slight reduction in the solar constant (or, for example, increase in global dust contamination) would be sufficient to initiate another ice age. In Figure 11 the global mean sea-level temperature, T_{sp}, is plotted as a function of the solar constant, S, currently assumed to be 2.00 ly min^{-1} (cf. Budyko used 1.92 ly min^{-1} as the current value !) The present value of T_{sp} is about 14.0°C. Three analyses were carried out, two with α allowed to vary according to $\alpha = b - 0.009\,T_o$ and $\alpha = b - 0.005\,T_o$, respectively, and the other with α held fixed at its assumed current values. There are no very dramatic events when α is kept fixed at the present values and S allowed to vary. However with a realistic albedo-temperature coupling a decrease of about 2% in the solar constant is sufficient to create another ice-age with ice-caps extending equatorward to 50° (i.e. where $T_o = -10$°C). Any further drop in the solar constant is accompanied in the model by a rapid transition to an ice-covered earth with albedo of 0.85 (cf. Equation (8a)) and an equilibrium temperature of -100°C. With the weaker dependence of α on surface temperature and ice-cover (i.e. 0.005 used in place of 0.009 and b adjusted accordingly) a severe ice-age would not occur until the solar constant decreased by about 5 - 6%.

Sellers also reports a few sensitivity studies where he has kept the solar constant fixed but has increased or decreased the effective diffusion coefficients in his model. He concludes that a relatively small change in the ability of the system to transfer heat poleward could conceivably offset any increase or decrease in the intensity of solar radiation. Hence the obvious caution about allowing only one variable to change without being able to allow for consequent changes in the others. Nevertheless Sellers' model implies quite conclusively that a decrease in the solar constant of less than 5% would be sufficient to start another ice-age.

3.4 The sensitivity of time-dependent Budyko-Sellers type models

Schneider and Gal-Chen (1973) examined the sensitivity of time-dependent models of the Budyko-Sellers type to changes in initial as well as external conditions. The problem is then formulated as an initial value problem in which the basic equations are integrated from a given initial state until a final asymptotic equilibrium state is obtained.

The time-dependent version of the zonally averaged, vertically integrated heat balance equation for the earth-atmosphere system is

$$C \frac{\partial T_0}{\partial t} = R + \operatorname{div} \underset{\sim}{F} \qquad (10)$$

where C is a prescribed thermal inertia coefficient for a zonal column through the land-ocean-atmosphere system. R is defined as in Equation (2) with I the net infra-red cooling to space given by

$$I = c(\varphi)\,\sigma T_0^4 \left[1 - m \tanh\left(19 T_0^6 \times 10^{-16}\right)\right] \qquad (11)$$

with $m = 0.5$. Note that with $c(\varphi) = 1$ (and ignoring the distinction between T_0 and T_s) this formula is identical to that used by Sellers (1969). Scheider and Gal-Chen claim to be using the zonally averaged 1000-mb temperature which has been written here as T_0, the notation used previously for the surface temperature. Their numerical scheme requires that the 'consistency factor' $c(\varphi)$ be adjustable over a small range of values close to unity.

Schneider and Gal-Chen examined four different versions of the model, defined by different formulations of $\operatorname{div} \underset{\sim}{F}$ and α.

(1) Labelled (S), is the formulation of Sellers (1969) with respect to $\operatorname{div} \underset{\sim}{F}$ and α.

(2) Labelled (SV), is exactly as (S) but with $\langle v \rangle \equiv 0$.

(3) Labelled (B), uses α as defined in (S) but has the Budyko (1969) formulation for $\operatorname{div} \underset{\sim}{F}$ as in Equation (4).

(4) Labelled (FG), is as (S) with respect to $\operatorname{div} \underset{\sim}{F}$ but uses a formulation for α attributed to Faegre (1972) such that

$$\alpha = 0.4860 - 0.0092\,(T_0 - 273)$$
$$0.25 \leq \alpha \leq 0.85 \qquad \text{for all values of } T_0$$

ie. α is a function of temperature everywhere.

The asymptotic equilibrium climate from the time-dependent version was shown to exhibit the same sensitivity to small changes in the solar constant as experienced with the steady-state models. Figure 12 shows the results of Schneider and Gal-Chen for their version (S) of the Sellers (1969) approach. Again with a decrease in solar constant of more than 1.6% a planetary mean

temperature of 174.65K was obtained corresponding to an entirely ice-covered earth. Calculations for a 1% decrease in solar constant and Sellers-type energy fluxes with, (S), or without, (SV), mean meridional motion gave quite similar results both in the global average amount of temperature decrease and in the latitudinal distribution of the temperature decrease. The results for the Budyko parametrization, (B), are qualitatively similar to those of (S) and (SV) for the global average. However the corresponding latitudinal distribution of temperature decrease differs significantly from that of Sellers. The particular form used for $d\omega E$ does appear to be significant for determining the details of the latitudinal distribution of temperature.

The stability of the (B) and (S) solutions to various perturbations of the initial temperatures were examined and the results are summarised in Table 5. It was found in both cases that the respective steady-state control climates could be recovered exactly if the temperature perturbation was positive. Perturbations of 2K and 16K were tried and in both energy-balance requirements alone produced a return to the control climate ie. the control climates appear to be unique, fully stable and transitive. For an initial temperature decrease of magnitude not exceeding 18.3K the apparent slight intransitivity in the equilibrium climates has been attributed to computer rounding errors. The models then show that the asymptotic equilibrium climate is extremely insensitive to changes in the initial conditions. However for initial temperature decreases greater than about -18.3K both models lead to an ice-covered earth solution similar to curve D in Figure 12. Thus it takes a very large negative perturbation to the initial temperatures to lead to the highly intransitive alternative ice-covered earth regime of these models. The insensitivity of the model to decreases in the initial temperature conditions of less than 18K follows from an unchanged albedo in equatorial latitudes.

Initial Temperature Perturbation (K)	Mean Planetary Temperature (K)	
	Sellers (S)	Budyko (B)
0	287.06	287.09
$0 < P < 16$ for all φ	287.06	287.09
$-18.3 < P \leq 0$ for all φ	286.67	286.87
$P < -18.3$ for all φ	175.58	175.44
-22 for $\lvert\varphi\rvert \geq 20^\circ$ 0 for $\lvert\varphi\rvert < 20^\circ$	286.67	286.87

Table 5. Asymptotic steady-state global average temperatures resulting from various perturbations to the initial temperature distributions for two of Schneider and Gal-Chen's (1973) parametrizations. This table is taken from the review of climate modelling by Schneider and Dickinson (1974).

The implication from these experiments is that even when ice-cover exists down to temperate latitudes but if the solar constant and the tropical albedos are unaltered then there is still sufficient energy available to the tropics to provide a poleward heat transport to melt the ice and restore the equilibrium climate to near the present-day value. If the equatorial temperature is reduced into the range of the positive feedback temperature-albedo coupling then climate instability is triggered, and an ice-covered earth results. This hypothesis was tested (see Table 5) by introducing a very large negative perturbation in the

initial temperatures, viz. -22K, for all $|\varphi| \geq 20^\circ$ only. In both models there was a return to near the present-day climate.

The model, (FG), which adopts Faegre's (1972) formulation for α is very sensitive to negative perturbations in both solar constant and initial conditions. The interpretation is that with the temperature-albedo coupling for all temperatures and at all latitudes then a slight negative perturbation in either initial temperature or solar constant can cause a decrease in energy input to the tropics thereby enhancing the albedo-temperature positive feedback at all latitudes. Experiments with such energy-balance models suggest that the stability of the climate may be quite dependent on the functional relationship between the albedo and the surface temperature.

3.5 Multiple solutions and their stability

We have seen from the numerical perturbation studies of Schneider and Gal-Chen (1973) that more than one solution is admissable in the Budyko-Sellers models with the present value of the solar constant ie. the system appears to be intransitive. North (1975) studied the latitude of the ice-line as a function of the solar constant by using a model, described as intermediate between that of Budyko and that of Sellers, which yielded exact analytical solutions. His results, summarised in Figure 13, show a multiplicity of solutions for the present-day value of the solar constant; indeed, there are no less than five solutions for the present value of the solar constant! Other studies with models of this type, but with different formulations for the individual terms, give results qualitatively similar to these. Lowering the solar constant by a few per cent leads to a catastrophic plunge to an ice-covered planet but more than a 30% increase is needed to yield an ice-free earth as the only solution.

North (1975) performed a linear stability analysis about the equilibrium solutions on branches I, II and III and confirmed that the present climate (I) and the ice-covered earth (III) solutions are stable (shown numerically by Schneider and Gal-Chen (1973)) while the intermediate solution (II) with the earth about two-thirds covered with ice is unstable and therefore of no physical significance.

Several other authors, including Held and Suarez (1974), Su and Hsieh (1976), Ghil (1976), Drazin and Griffel (1977), North (1977) and Cahalan and North (1978), have established a general stability theorem applicable to a wide variety of specific models within the general class. The slope stability theorem states that

$$\frac{dx_s}{ds} > 0 \implies \text{Stability},$$

$$\frac{dx_s}{ds} < 0 \implies \text{Instability},$$

where x_s is the sine of the latitude of the ice-line.
The validity of this theorem for a large class of models is important, since it has implications for those branches of the solutions nearest to that corresponding to the present climate, viz. they are unstable and therefore unphysical. Their instability also explains why they were not located by the earlier approaches of Budyko, Sellers and Schneider and Gal-Chen.

One interesting point is that the slope stability theorem implies that if a cusp near $x_s = 1$ exists in a model solution (as, for example, in Figure 13) then small ice-caps are unstable. However the existence of such a cusp appears to depend very crucially on the details of the parametrizations

adopted. Budyko-Sellers type models may be too simple to examine such a point. This is unfortunate since whether small ice-caps are stable or not is an important issue if one recalls the concern over rising levels of CO_2 in the atmosphere and the possible effects of melting ice on the sea-level.

3.6 Further studies with Budyko-Sellers type models

The basic concept of the Budyko-Sellers type models has been extended and made more realistic and much more complicated in a great variety of ways. The comprehensive review by Schneider and Dickinson (1974) and the more recent reviews by North (1979), North et al (1980) and Hunt (1980) are recommended as sources for further study.

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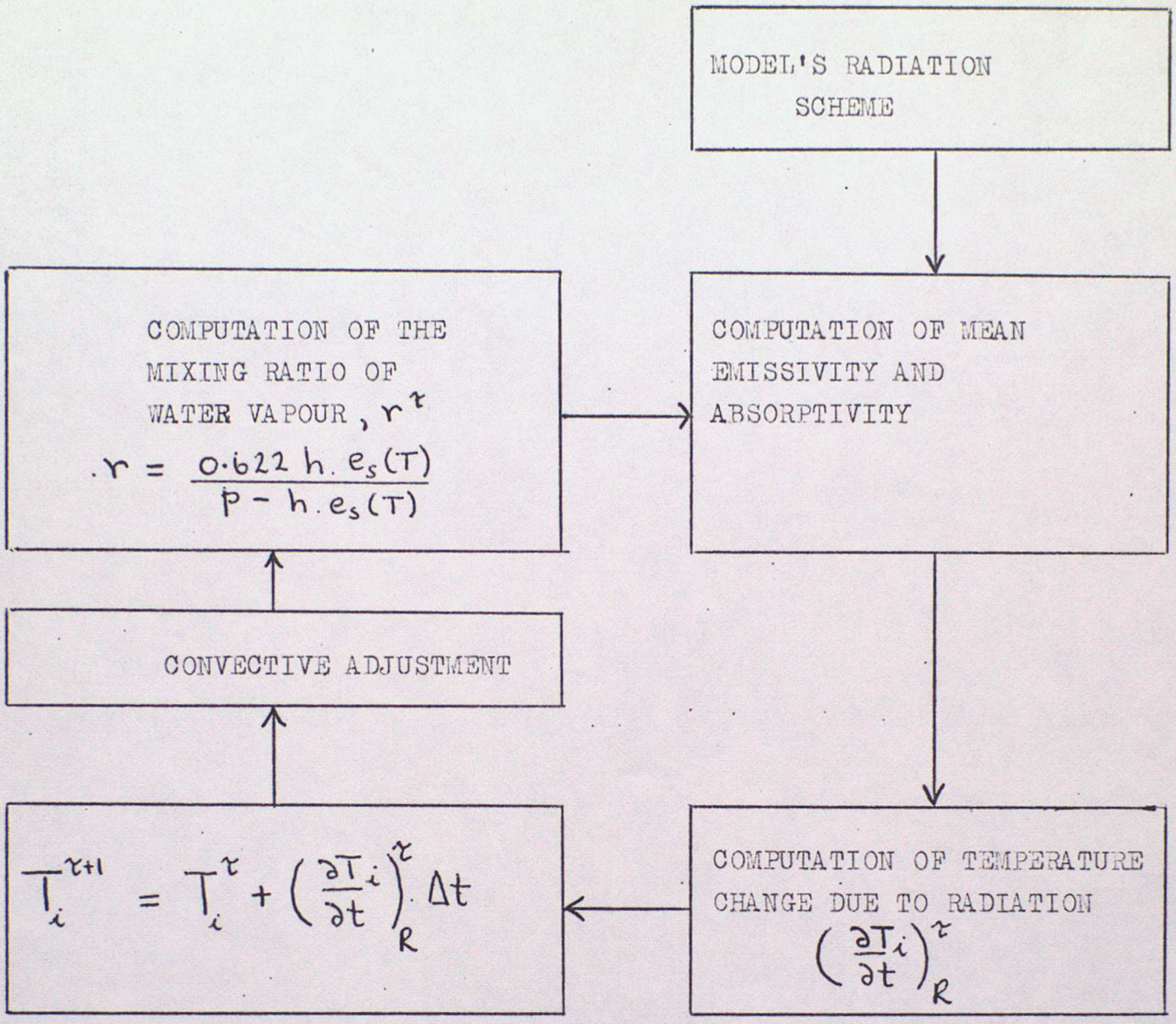

Figure 1. Schematic representation of the iterative procedure used to obtain the state of radiative-convective equilibrium, as described by Manabe and Wetherald (1967).
τ is the number of the time-step of the numerical integration.
i is the index for the finite-differencing in the vertical.

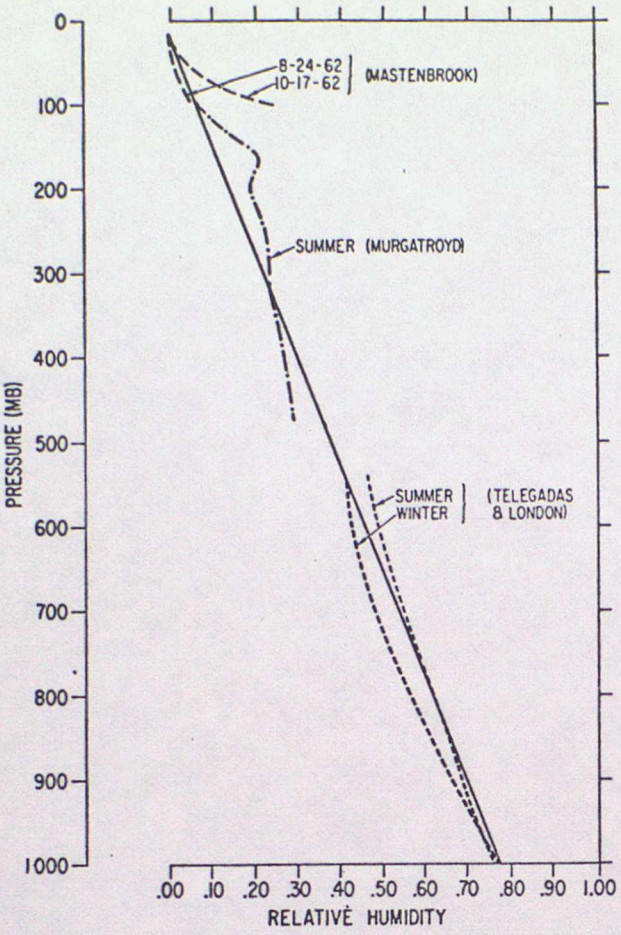

FIG. 2 Vertical distribution of relative humidity (Mastenbrook, 1963; Murgatroyd, 1960; Telegadas and London, 1954).

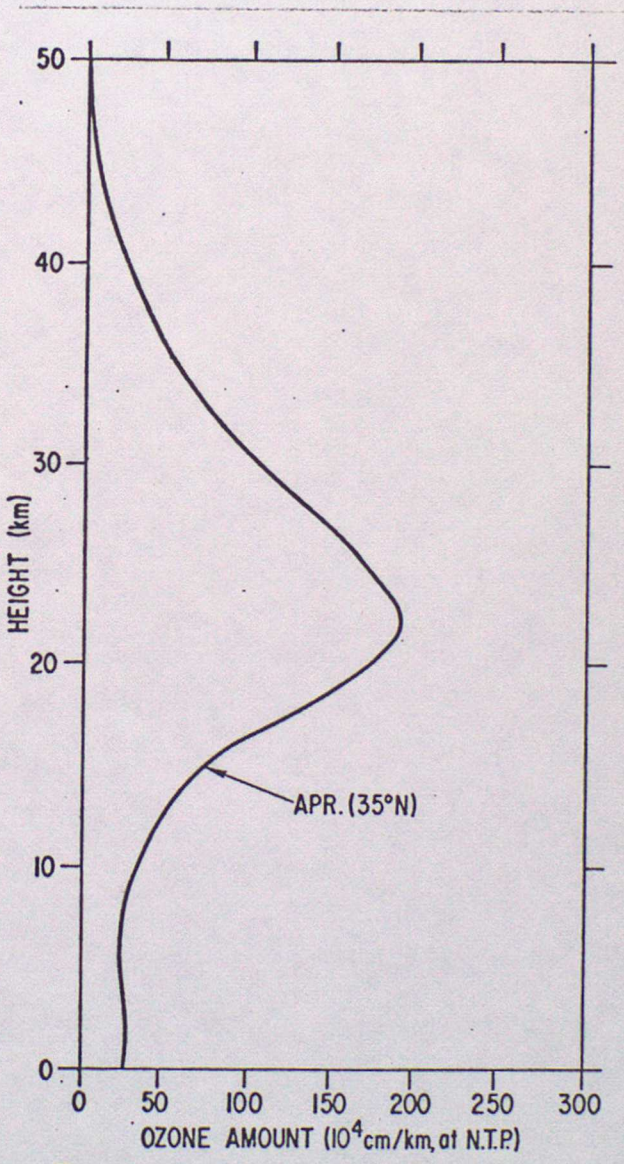

FIG. 3. Vertical distribution of ozone at 35N, April (Herring and Borden, 1965), normalized by the total amount from London (1962).

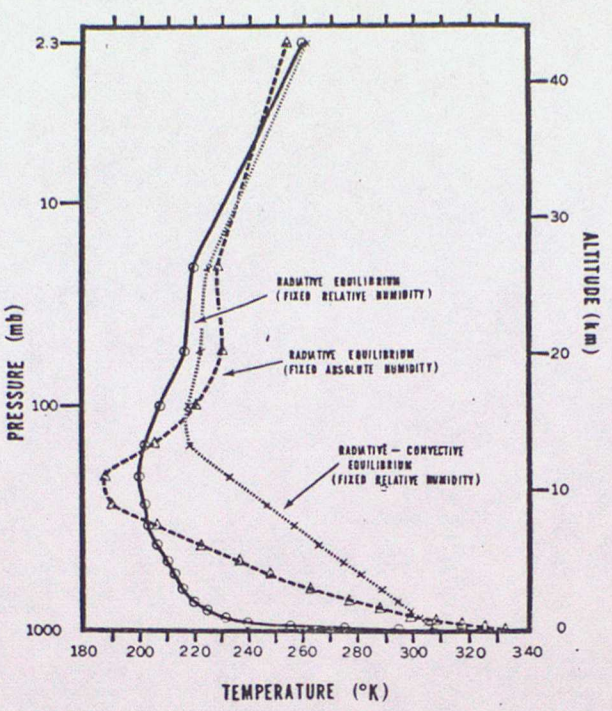

FIG. 4. Solid line, radiative equilibrium of the clear atmosphere with the given distribution of relative humidity; dashed line, radiative equilibrium of the clear atmosphere with the given distribution of absolute humidity; dotted line, radiative convective equilibrium of the atmosphere with the given distribution of relative humidity.

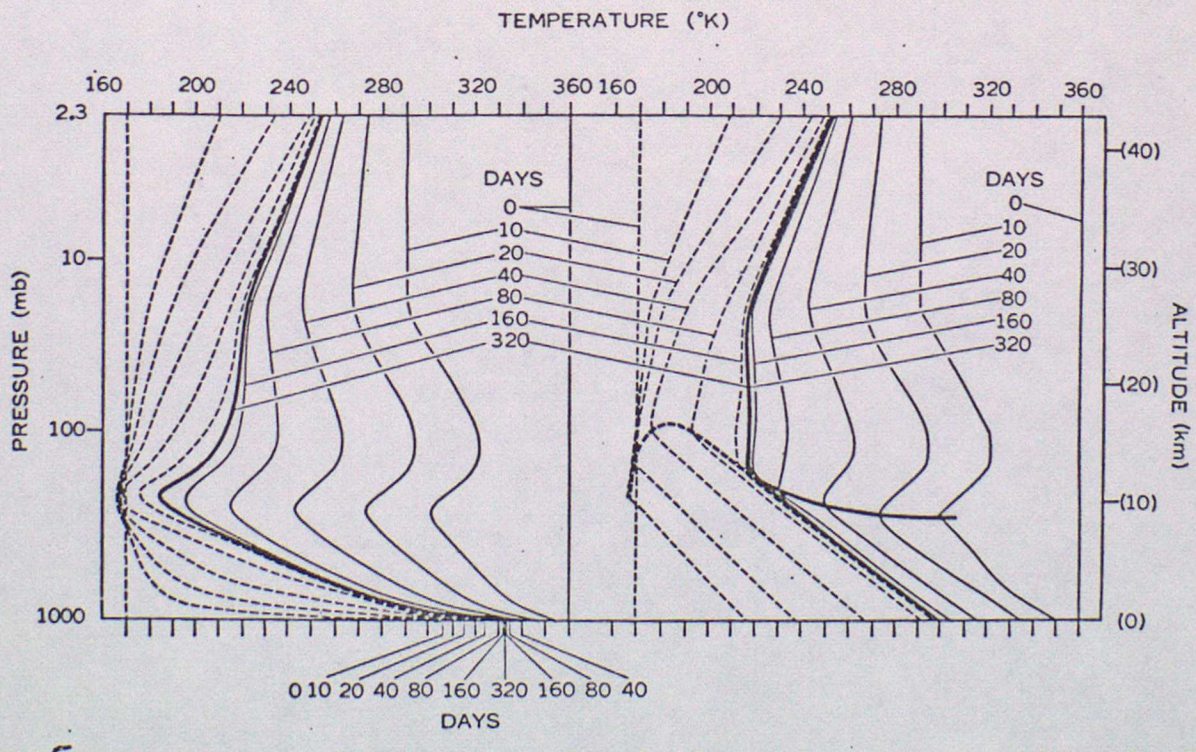

FIG. 5 The left and right hand sides of the figure, respectively, show the approach to states of pure radiative and thermal equilibrium. The solid and dashed lines show the approach from a warm and cold isothermal atmosphere.

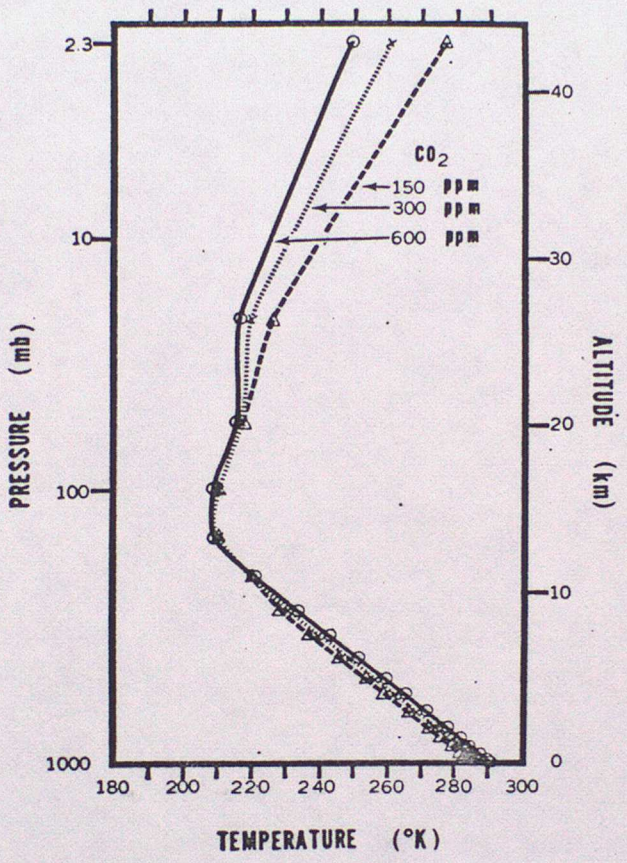

FIG. 6. Vertical distributions of temperature in radiative convective equilibrium for various values of CO_2 content.

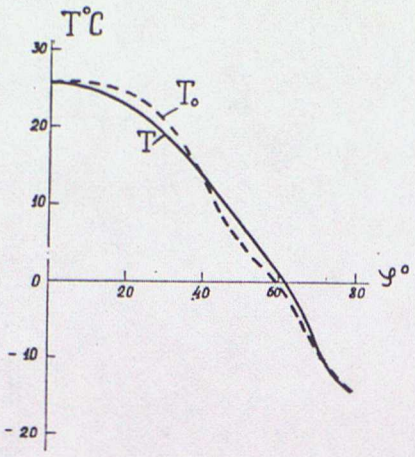

Fig. 7. The average latitudinal temperature distribution.

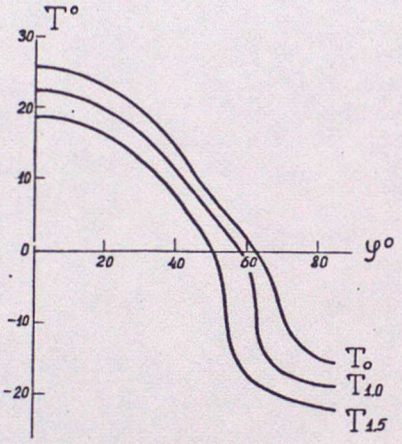

Fig. 8. The dependence of temperature distribution on radiation amount.

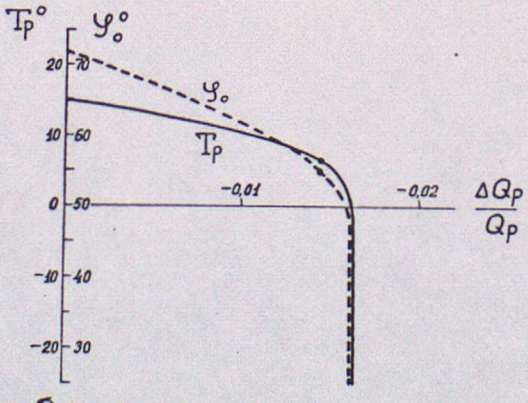

Fig. 9 The dependence of the Earth's temperature and ice cover boundary on radiation variations.

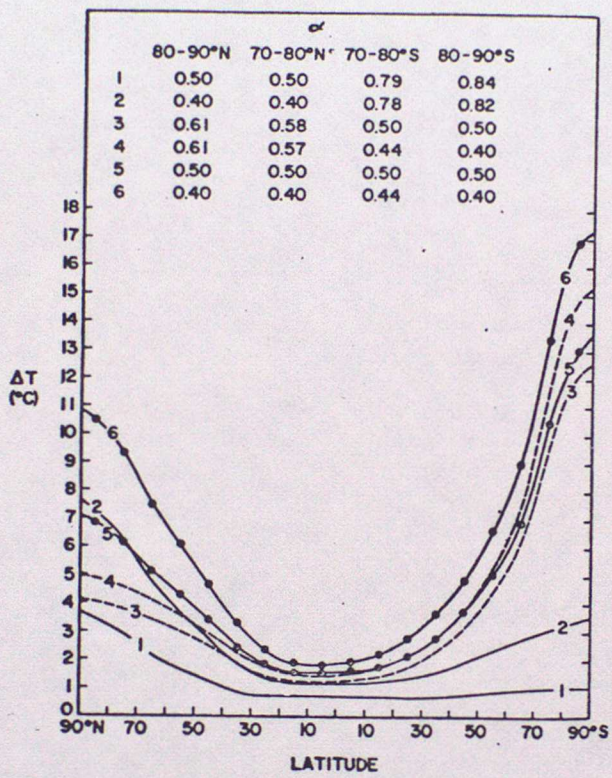

FIG. 10 Predicted latitudinal distribution of the mean annual temperature rise associated with albedo manipulation at one or both poles.

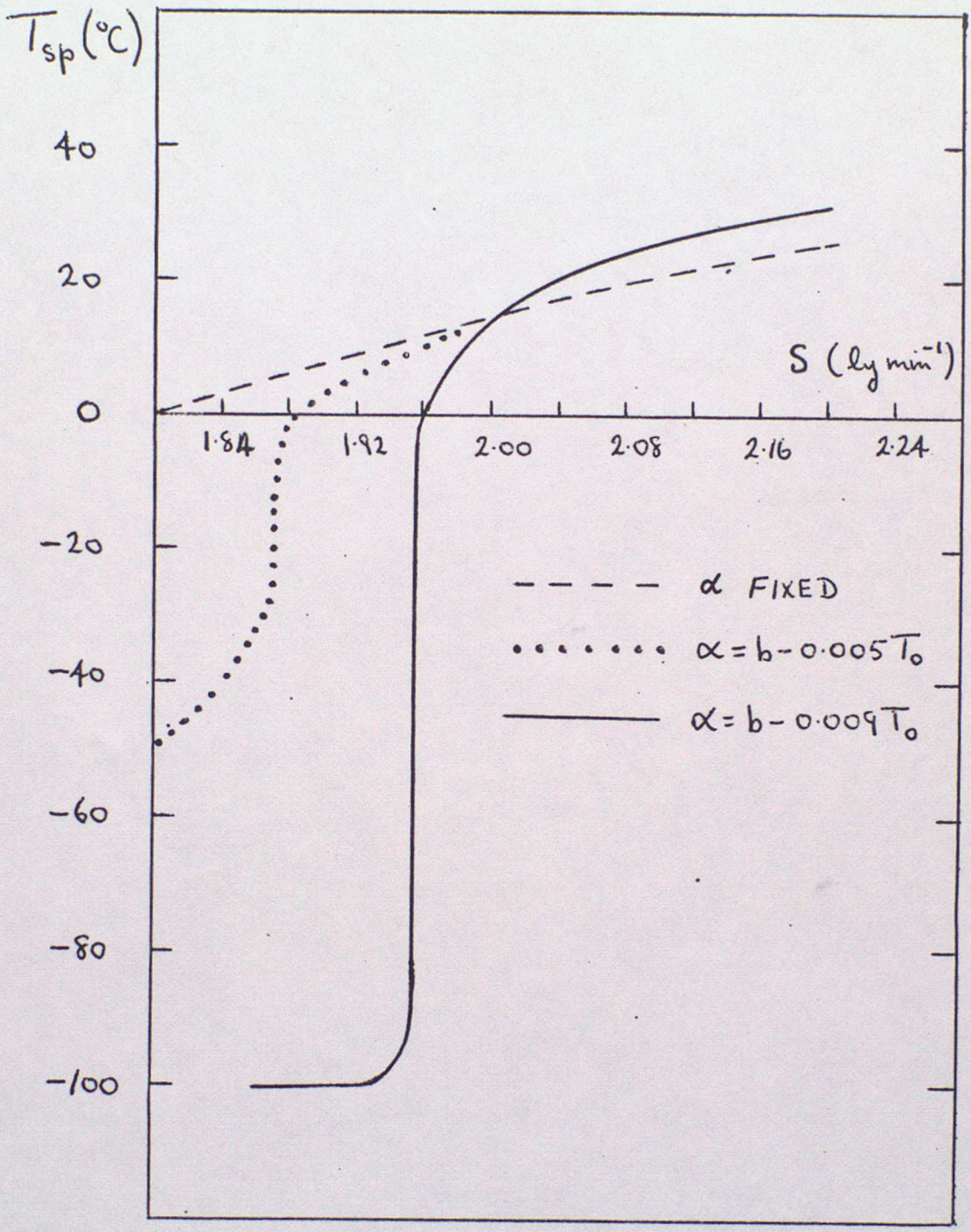

Fig. 11. The mean global sea level temperature, T_{sp}, as a function of the solar constant, S, for different specifications of the planetary albedo, α. This figure is based on the results of Sellers(1969).

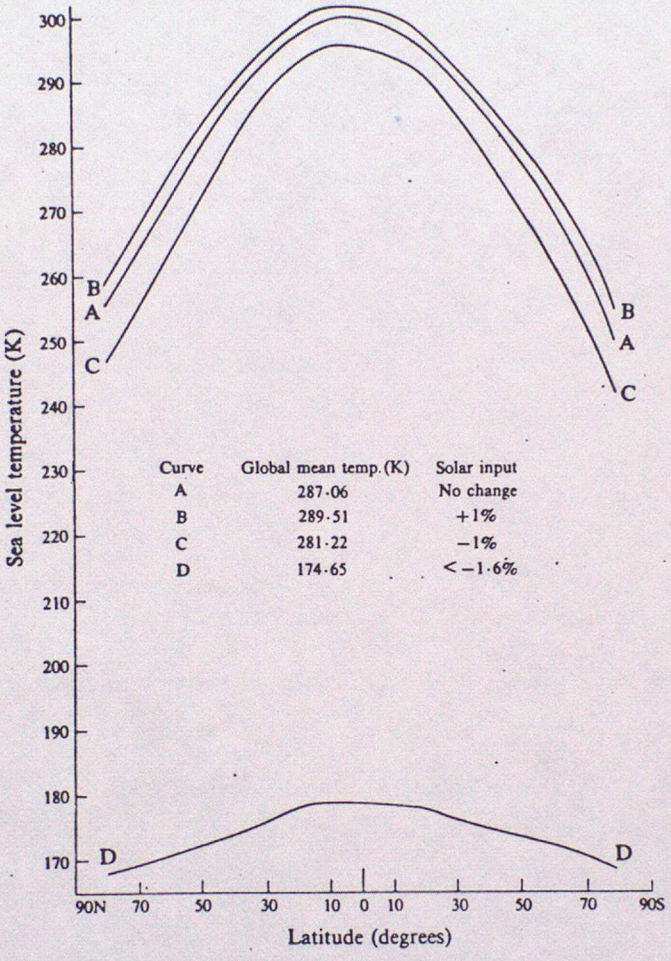

Fig. 12 Latitudinal temperature distribution calculated for various solar constant values. [From Schneider and Gal-Chen (1973).]

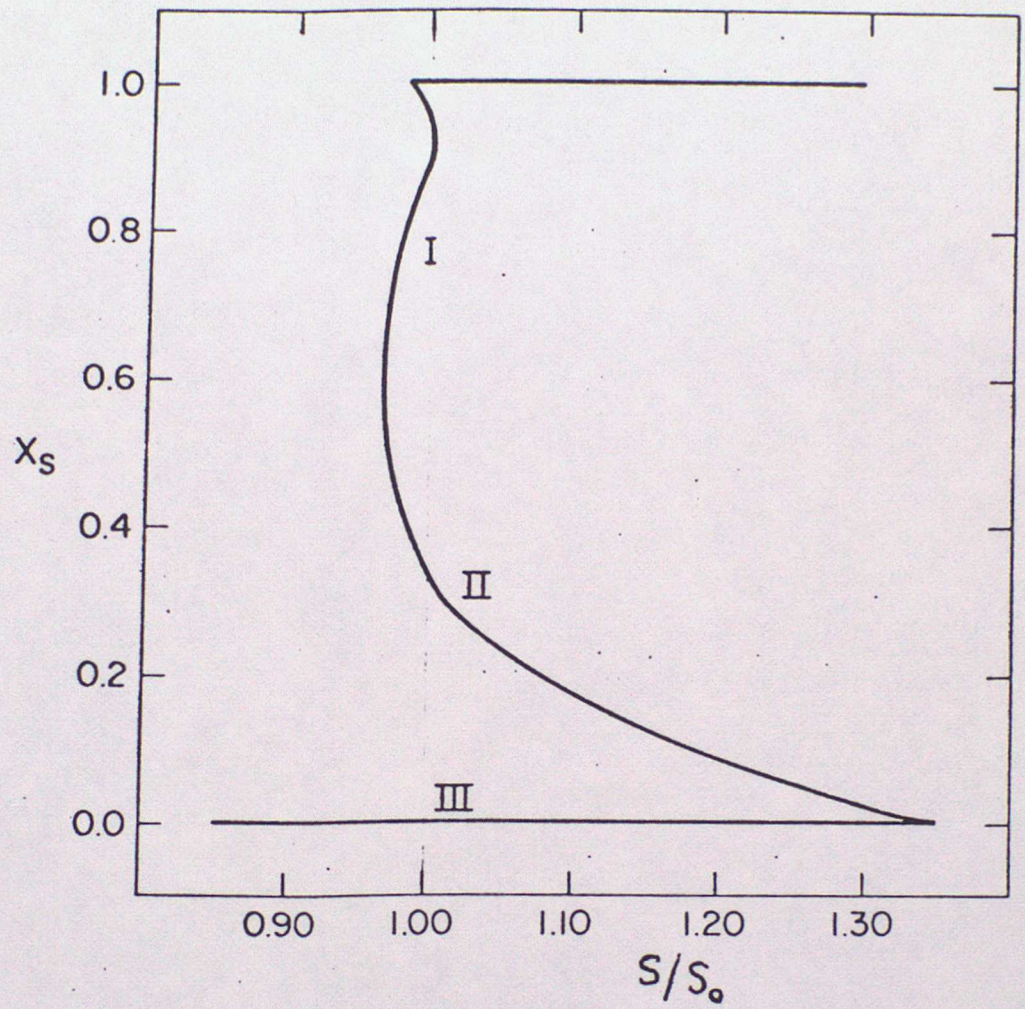

Fig. 13 A typical plot of the sine of the latitude of the ice line (X_s) versus the solar constant S/S_0 normalized to its present value. This curve is for an exact solution of a model with constant diffusion coefficient and discontinuous albedo at the ice line.

This figure is taken from North(1975).

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CLIMATIC FLUCTUATIONS AND TRENDS IN THE INSTRUMENTAL PERIOD

D.E. Parker Met O 13

Advanced Lectures 1982

ABSTRACT

The reality and significance of fluctuations in climate in the last 300 years are discussed in the light of the quality and quantity and homogeneity of the data. Possible causes of recent climatic fluctuations are presented. Indications from upper air observations, covering the last 30 years, are included.

1. Introduction

This presentation covers climatic flucations on time-scales from a decade to a few centuries, and bridges the gap between interannual variability and paleoclimatology, which have been covered in preceding lectures. The approach to climatic fluctuations in the instrumental period has been traditionally somewhat different from the approach to interannual variability, with emphasis on a few long well-documented local or global time-series of data, rather than on the stistical and physical links between atmospheric behaviour in different parts of the globe. Nevertheless, interannual and climatic fluctuations are part of the variability of the same physical system; short-term and longer-term aspects may be closely linked; and there is no formal dividing line between synoptic climatology and studies of climatic change.

2. Observations of climatic fluctuations of temperature

a) Local time-series

The fundamental aim in constructing local time-series is to ensure that the data are as temporally homogeneous as possible: in other words, that a given meteorological reality has always resulted in the same value in the series. Only to the extent that this is the case, can the series be used to make reliable deductions about climatic fluctuations.

Manley (1974) constructed a monthly temperature series for Central England, and a smoothed seasonal series derived from this is shown in Figure 1. Manley chose to represent an area, rather than to reduce his series to a single point, because of the scattered locations of most of the early observations, and because the only concentrated observations were in London and thus were affected by heating resulting from urban expansion. He adjusted the monthly data to compensate for the following variables:

i) Times of observation: the traditional mean temperature computed as $\frac{1}{2}$ (max + min) differs from eg the mean of 8 am and noon, by amounts which vary with season but can be estimated from modern data.

ii) The change of calendar in 1752: daily data facilitated this adjustment.

iii) Housing of thermometers: in the 18th Century this was often an unheated room attached to a house; and in later years use was often made of either north-wall observations, or of Glaisher screens, which preceded Stevenson screens.

iv) Types of thermometer: pressure corrections were needed for early thermometers which consisted of a U-tube with one open end. Also different scales had to be converted to modern units. Unreliable instruments were excluded from Manley's compilation.

v) Geographical region and site: Manley found from modern data that, so long as frost-hollows and other unrepresentative sites are excluded, the monthly deviations from a normal, or the successive differences between months of the same name, are almost uniform for English inland stations. Thus he could work with differences rather than actual values, and produce single series from overlapping records at sites scattered through much of England. This technique also aided the elimination of uncertainty in the temperature scales and in the effects of the observing time and instrumentation.

Manley checked his compilation against the series for Utrecht in Holland, and found good agreement. Also, in periods of sparse data he took note of observations of wind direction and snowfall.

It can be seen from Manley's series, shown in Figure 1, that, except in summer, the 20th Century in Central England has been warmer than the previous 250 years: the peak warmth was in the mid-20th Century except in winter when it was earlier. The late 17th Century was particularly cold and is often regarded as the peak of the "Little Ice Age" of the 16th to 19th Centuries. There was notable temporary warmth in the 1730s. These features are also found in temperature series for Utrecht and for other locations in Northwest Europe.

The possible effect of urban heating, avoided by Manley, is illustrated for Paris in Figure 2 (after Dettwiller 1978). Paris has warmed by 1°C since 1900 in comparison with the surrounding countryside. Studies such as Dettwiller's can be used to compensate urban data only if there are adequate nearby rural observations.

b) Global or hemispheric series

i) Surface

The mean surface temperature of the globe must in practice be approximated by the mean for a selected number of observing points or, at most, for a selection of regions. This applies even at present, as parts of the oceans are still incompletely observed; but was emphatically the case in the early instrumental period when temperature series were limited to Northwest Europe, parts of North America, and Japan. Nevertheless, early global or hemispheric series derived from such scattered observations have a major point in their favour. This is that the observations were mainly over midlatitude land areas, where the variability of temperature is greater than elsewhere: greater than over the oceans because of the smaller effective thermal capacity of the surface, and greater than in the tropics because of the combination of greater mean north-south temperature gradients and more variable meridional circulation in midlatitudes. Thus the series, being dominated by the most variable parts of the globe, will be highly correlated with the true global mean, though the variance may be too great. A possible disadvantage is that, if the true mean temperature remained constant but the phases of the midlatitude long waves changed, the mean temperature computed from the station data could have changed because the

geographical distribution of Europe, USA and Japan accords rather well with a wave pattern with wavenumber 3. The possibility can even be entertained, that the "Little Ice Age", although well-documented in Europe and to some extent in USA and Japan, was, so far as a hemispheric average is concerned, an illusion based on such a sampling of the long waves.

For more recent times, data have covered much of the continents and the sampling problem will have been reduced. Borzenkova et al (1976) used all available data from 1881 to compute annual temperature anomalies over a set of regular grid points at 20°N to 90°N (Figure 3). The station data were first reduced to deviations from the normal, and comparisons of such deviations with those for neighbouring stations were used to eliminate suspect data. Absolute values cannot be used in this way because of the effects of altitude and siting differences. Borzenkova et al also used a limited amount of data over the oceans. Figure 3 indicates that the mid 20th Century warmth was a hemispheric as well as European phenomenon.

In view of the geographical distribution of temperature variability, Landsberg et al (1978) were able to select nine stations to produce an acceptable linear regression to approximate Borzenkova et al's series. Figure 3 also includes Landsberg et al's computed series. However the occasional poorly simulated years (eg 1922) illustrate the weakness of omitting the oceans and subtropics, which were included by Borzenkova et al.

Groveman and Landsberg (1979) used regressions relating Borzenkova et al's series to values at a few very long-period stations, to produce a Northern Hemisphere temperature series back to 1579 (Figure 4). Different segments of the series were based on different regressions (always explaining over half the variance) according to the data available. Features in common with European series such as Manley's, are coldness in the late 17th Century and warmth around the 1830's. The early 17th Century was also cold in Europe (Lamb 1977).

As work progresses in homogenizing and analysing early instrumental data, it is hoped to reduce sampling and other uncertainties associated with time-series such as that in Figure 4 and to clarify the nature of the "Little Ice Age".

ii) Upper-air

It is only for the most recent decades that upper-air data have been available for monitoring climate. As with surface data, it is necessary to compensate for changes in instrumentation. Figure 5 shows that 1978-80 sonde temperatures over Crawley in summer were lower than expected on the basis of surface temperatures, because the Kew MkIIb sondes had been replaced by MkIII. Corrections applied from 1981 at the station appear to have compensated for the change of sonde. It is also necessary to avoid biased sampling by using a regular upper-air network, dense in comparison with the scale of the atmospheric long waves. Angell and Korshover (1978) used as regular a network as possible to monitor world-wide surface and upper-air temperatures for 1958-77. Their results (Figure 6) show initial cooling followed by a regime of fairly constant temperatures. A cool period appears to have followed

the eruption of Mount Agung in Indonesia in March 1963. However, the network of upper-air stations used was rather sparse in the tropics. On the regional scale, initial results for a dense Arctic network of stations suggest a tendency to warming in recent years (Figure 7).

Other authors (e.g. Painting 1977) have tried to obtain upper-air data with an even coverage by extracting values at grid-points from archived series of operational analyses. But this has the disadvantage that the grid-point data obtained from different national centres disagree, often substantially, especially over the oceans (Figure 8) where data are sparse and operational techniques of forecasting and analysis give a variety of results. Furthermore, analysis methods and data availability change with time, not necessarily in the same manner at different national centres, so that the grid-point data from one centre may show spurious climatic changes relative to those from another centre. Ultimately it is hoped to fill the oceanic gaps by means of satellite radiance soundings: considerable development is in progress at present with the aims of improving the satellite sampling in cloudy areas and of developing satellite instrumentation of lasting quality.

3. Circulation changes and atmosphere-ocean interaction

The discussion of the reality on a global scale of the "Little Ice Age" has already included consideration of circulation changes, in terms of changes of phases of the long waves. Lamb (1977) supports the hypothesis that such changes have occurred, mainly in the form of eastward progression in the early 19th Century (Figure 9). His results are based on a careful compilation of surface pressure and wind data over Europe, the Atlantic and North America.

On shorter time-scales, Van Loon and Williams (1976a, b) have studied the relationship between circulation trends and temperature trends in the present century. The reader is referred to their papers which contain complex charts comparing the trends of the two parameters. In general their results are as expected: for example, a trend to increasing easterly flow over Europe in winter was accompanied by a cooling trend. However their study was geographically limited by lack of data over the oceans, Arctic and tropics.

An index of Atlantic circulation, shown in Figure 10, is the strength or location of the westerlies as indicated by the difference in mean-sea-level pressure between the Azores and Iceland. The long-term trend in this index in winter roughly parallels the trend of Manley's Central England Temperature as expected, with greatest warmth paralleling the strongest westerlies earlier in this century, but correspondence is poorer in the late 19th Centry.

The relationship betweenlocal temperature and regional circulation is amply demonstrated by Van Loon and Williams' papers and by comparison of Figures 1 and 10. In addition it is possible that a change of circulation on the large scale may influence the mean global air temperature because local air temperature changes may not cancel. One physical mechanism for this would be changes in cloud amount, type and geographical distribution, all of which would affect the Earth's radiation balance. See for example Potter et al (1981). A further mechanism could be re-partitioning of heat between atmosphere and ocean, because heat fluxes between sea and air depend on the air temperature and velocity, which would change if the circulation changed, and the changes in heat fluxes might not cancel for the ocean as a whole. Unfortunately it is difficult to check this hypothesis because resulting oceanic temperature changes would be small, because the changes in oceanic heat

content would be distributed throughout the 50 to 100 metres of mixing layer which has a thermal capacity an order of magnitude greater than that of the atmosphere. It is easier to demonstrate that the surface (as opposed to mixing layer) temperature of the ocean influences the mean air temperature over large regions of the globe but on shorter (seasonal) time-scales (Navato et al 1981). Also a comparison of European temperatures (Figure 1) with North Atlantic sea surface temperatures (Figure 11) reveals some parallel tendencies over as long as the last century. Note that care is needed in constructing time-series of sea-surface temperature: instrumental corrections of up to 0.5°C may need to be applied (Caton and Lawes 1981), and the effective depth at which the water temperature is measured is important because of vertical temperature gradients.

4. Almost-intransitivity

Lorenz (1976) suggested that the climate might of its own accord be able to fluctuate between two or more stable states without any change of external forcing such as solar radiation. He described such a characteristic as "almost-intransitivity". If the climate were, on the other hand, transitive, it would only be able to display a single set of statistics for a given external forcing; and if it were totally intransitive its statistics for given external forcing would be determined by the initial conditions, and it would not of its own accord change from one set of statistics to another. Lorenz illustrated almost-intransitivity in the abstract by means of an interation of an algebraic expression which displayed long but irregular alternating periods of positive and negative values. For the real atmosphere, Figure 12 may suggest almost-intransitivity: at Dakar, Senegal, annual rainfall appears to have been persistently low around 1896-1903, 1910-14, 1919-26, and especially 1970-81, which period began with the well-known Sahel drought.

5. Radiative influences on climate

There are several effects which may have caused climatic changes in the instrumental period by altering the Earth's radiation balance. Chiefly investigated have been the man-made increase of carbon dioxide, which absorbs outgoing long-wave radiation; volcanic eruptions, the dust from which, if it reaches the stratosphere, where it can persist for several years without being washed out by rain, affects particularly incoming but also outgoing radiation; and changes of solar constant, usually in terms of sunspot number, whose relationship with solar constant is uncertain. All three were investigated by Miles and Gildersleeves (1977, 1978). When they studied the past 100 years, (Figure 13), they found significant statistical relationships between Northern hemispheric temperature series and the carbon dioxide concentration and volcanic dust, the latter being associated with lower temperatures: but there was no relationship between the temperature series and sunspot number. The relationship with volcanic dust did not hold for the longer period from 1665 (for which Central England Temperature was used as an approximation to hemispheric temperature); but Taylor et al (1980) studied specific major volcanic events since 1800 and, by compositing the results, found a suggestion of reduced global temperatures in the two years following eruptions. An additional radiative cause of climatic changes in historical times may have been man-made changes of the albedo of the earth's surface by deforestation and land development (Sagan et al 1979).

The greatest research effort at present is into the effects of increasing atmospheric carbon dioxide, the future increases of which have been estimated for the next 50 years under a variety of assumptions about future economic development and about the proportion of carbon dioxide which the oceans will

absorb (Pearman 1980). The estimates of oceanic absorption, which vary from 25% to over 50%, are based on observed changes in atmospheric carbon dioxide concentration since about 1900, and on estimates of the carbon dioxide output from industrial development and changes of land use. The consensus of predictions is for a doubling of atmospheric carbon dioxide in about 50 years. Numerical model integrations indicate a global average warning of about 2°C in equilibrium for doubled atmospheric carbon dioxide concentration, with greater warming at high latitudes. However, because of the large thermal capacity of the oceans and the slow circulation of their deep waters, there may be a time-lag of decades before equilibrium is reached. There is clearly an increased need to monitor the present climate carefully to detect any changes and investigate their causes.

6. Conclusion

The complexities of the study of recent climatic fluctuations have been demonstrated. Much work is necessary in this field in order to establish the causes of such fluctatuions with more confidence. The importance of climatic fluctuations for human life hardly needs to be stressed: for us in the West it is a matter of occasional inconvenience, but for some parts of the world, a matter of life or death (Figure 12). Therefore, worldwide co-ordinated research should be encouraged until these problems are much better understood.

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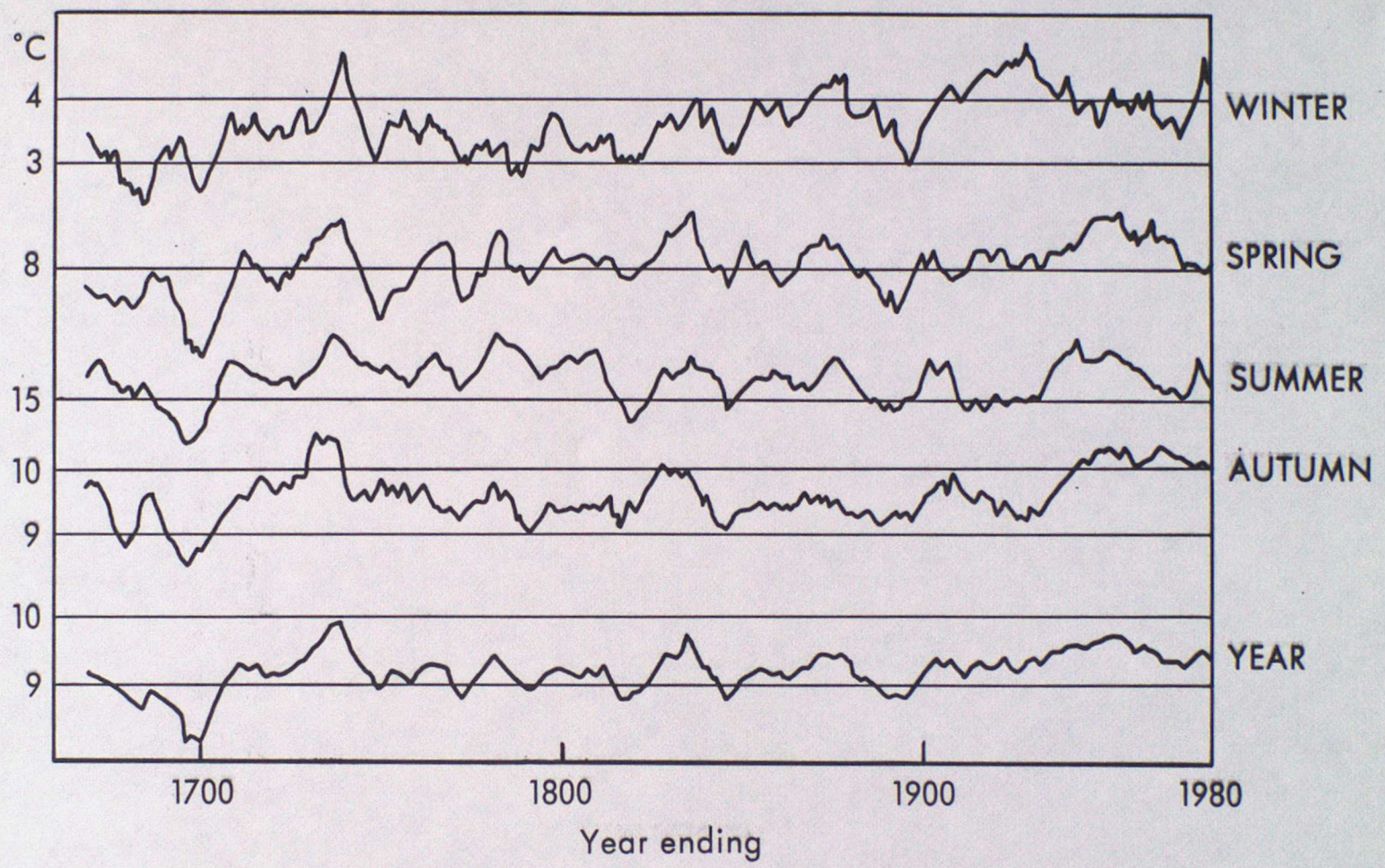


FIGURE 1. SEASONAL AND ANNUAL CENTRAL ENGLAND TEMPERATURES (DATA UP TO 1973 FROM MANLEY 1974). Values are decadal running means.

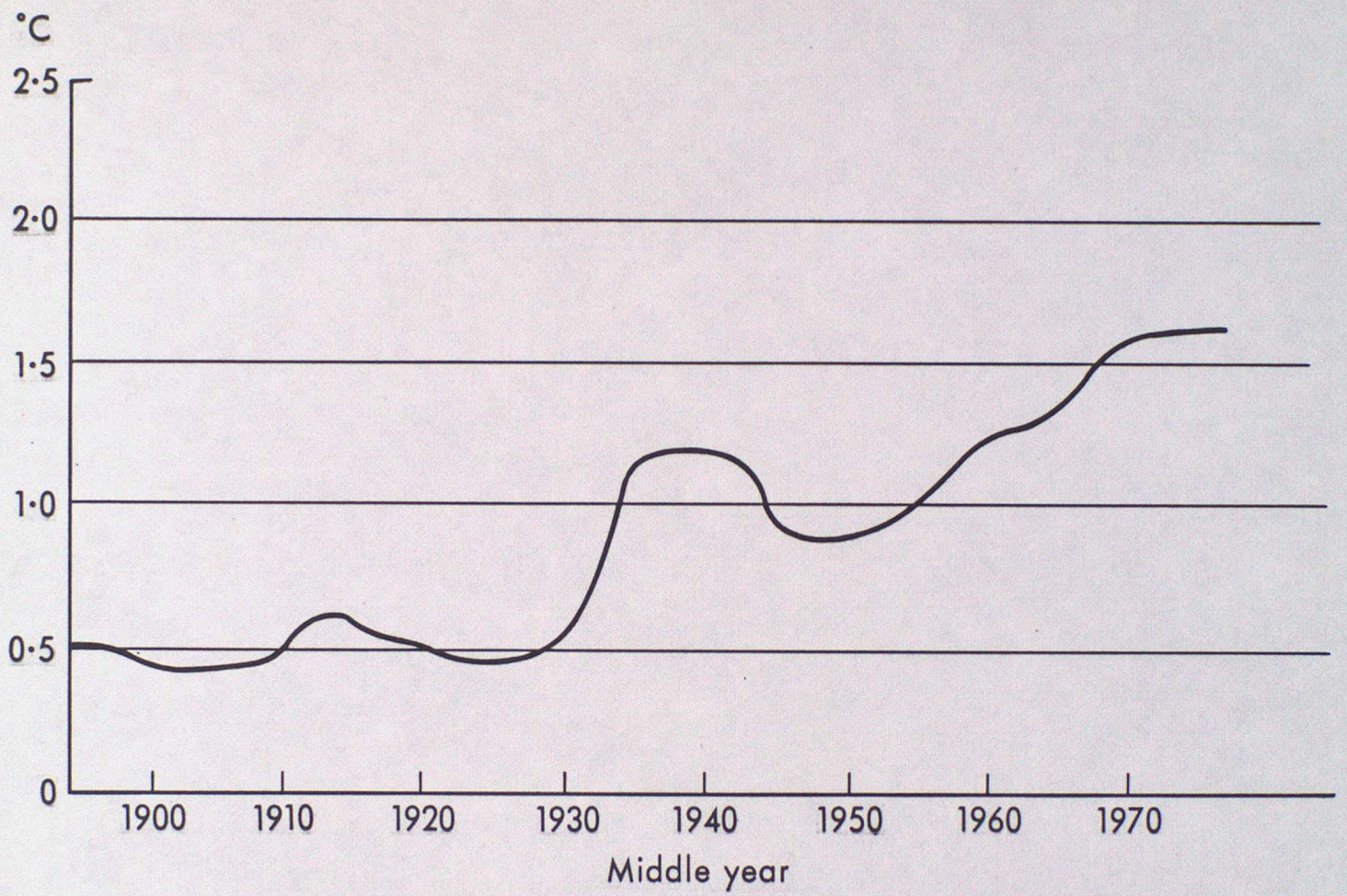
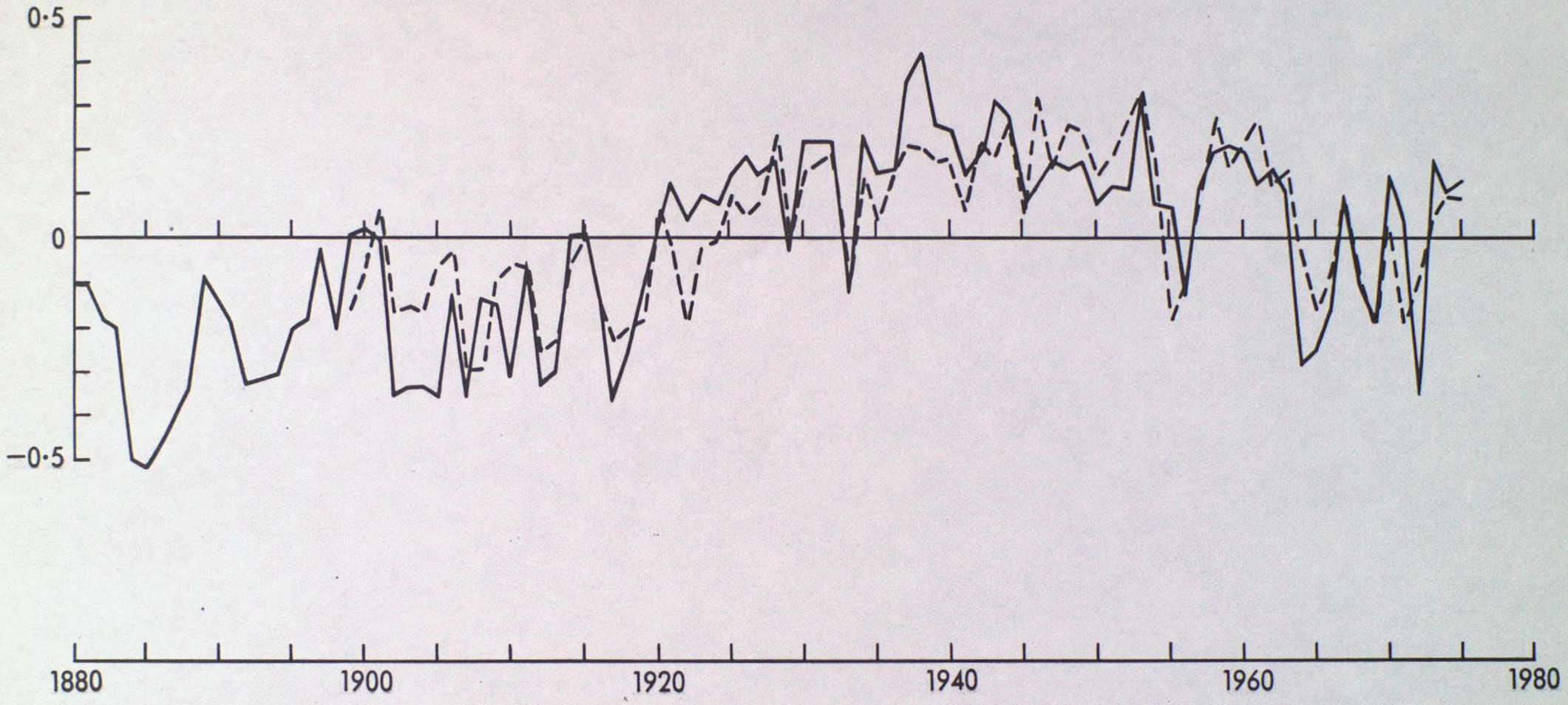


FIGURE 2. ANNUAL MEAN TEMPERATURE DIFFERENCE, PARIS—MONTSOURIS MINUS BESANCON (AFTER DETTWILLER 1978). Values are 10-year running means.



Solid line based on data by Borzenkova et al., (1976);
Dashed line simulation model of Landsberg et al., (1978)...

FIGURE 3. NORTHERN HEMISPHERE TEMPERATURE DEVIATIONS 1881-1975

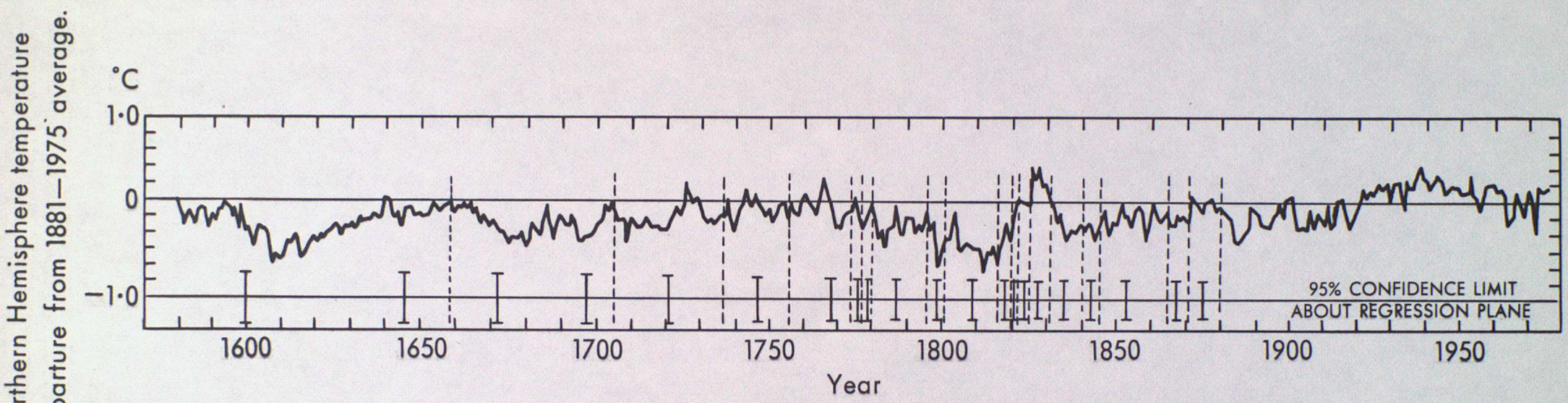


FIGURE 4. RECONSTRUCTED ANNUAL TEMPERATURE DEPARTURES (1579—1880) FOR THE NORTHERN HEMISPHERE FOLLOWED BY DATA OF BORZENKOVA et al., (1976). AFTER GROVEMAN AND LANDSBERG (1979). Confidence limits of regressions at bottom of graph.

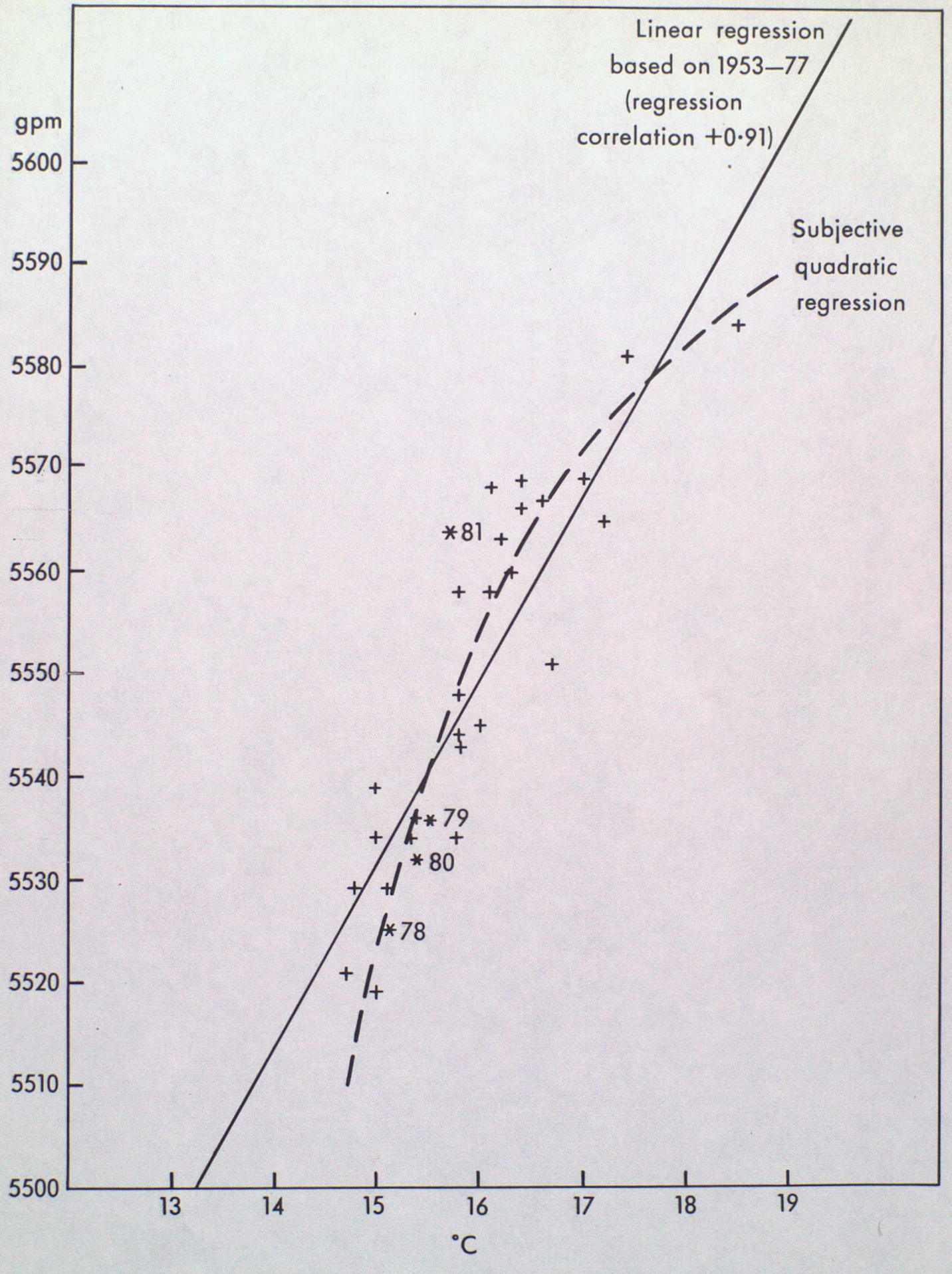


FIGURE 5. CRAWLEY 1000—500 MB THICKNESS VERSES S. AND SE. ENGLAND REGIONAL TEMPERATURE, SUMMER 1953—81

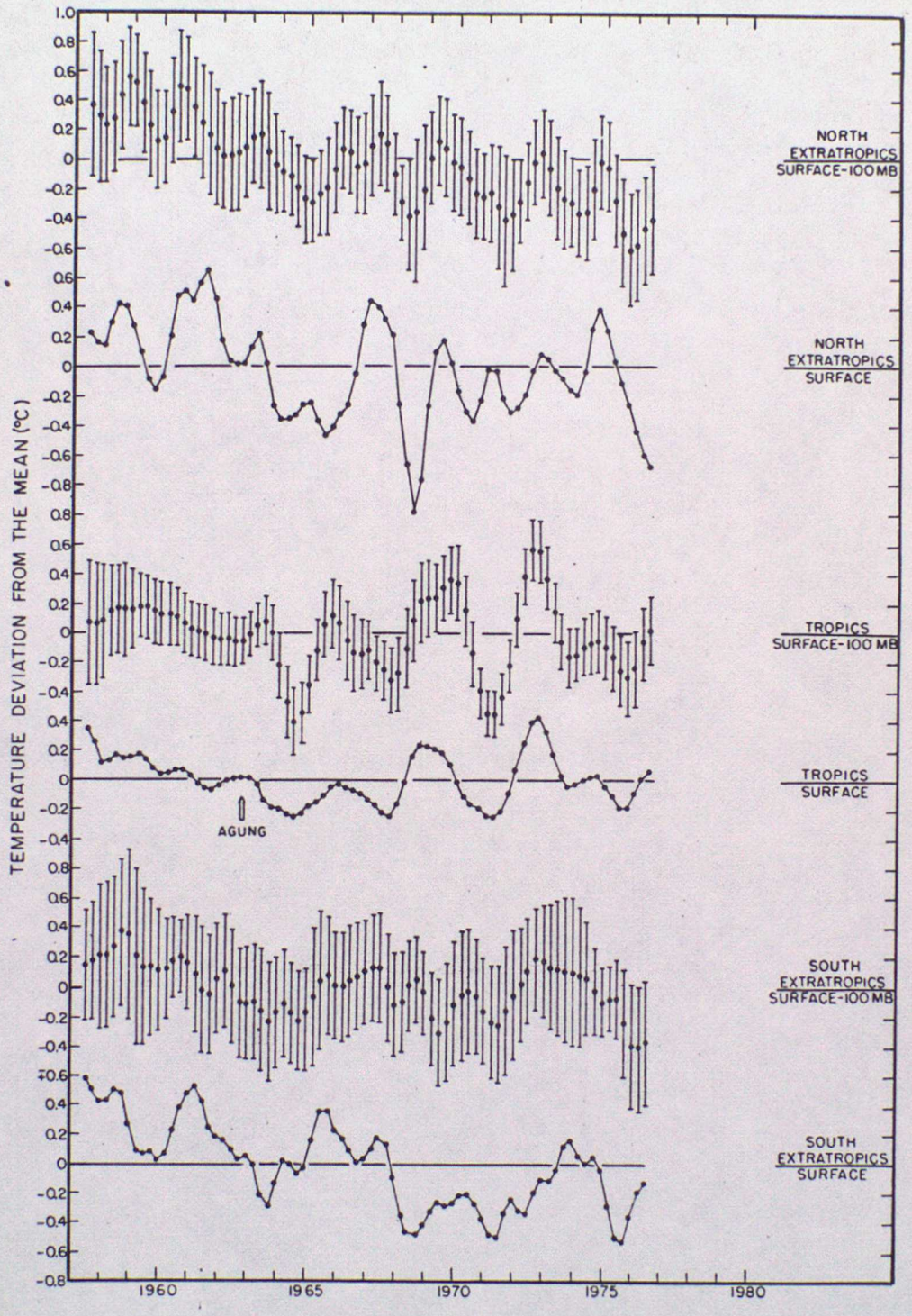
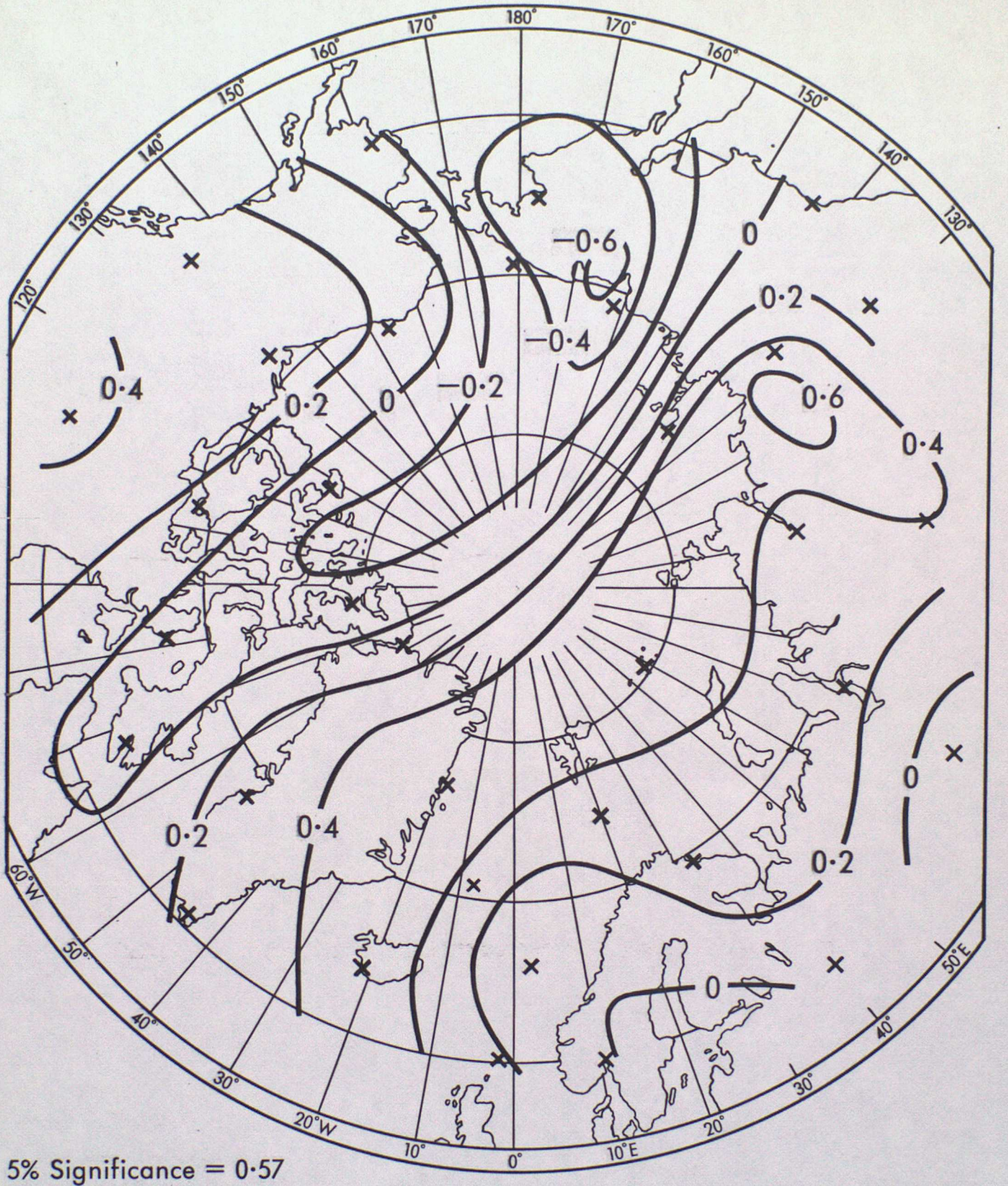


FIGURE 6. RECENT TEMPERATURE TRENDS DEDUCED FROM SURFACE AND UPPER-AIR DATA (AFTER ANGELL AND KORSHOVER 1978). The time-series and the ±2 standard deviations confidence limits have been twice smoothed 1–2–1 (1–1 at beginning and end). The tick marks are in summer of the given year.



5% Significance = 0·57

× = Upper-air station used in the analysis.

FIGURE 7. LINEAR TREND CORRELATION OF 1966—78 SPRING 1000—500 MB THICKNESS.

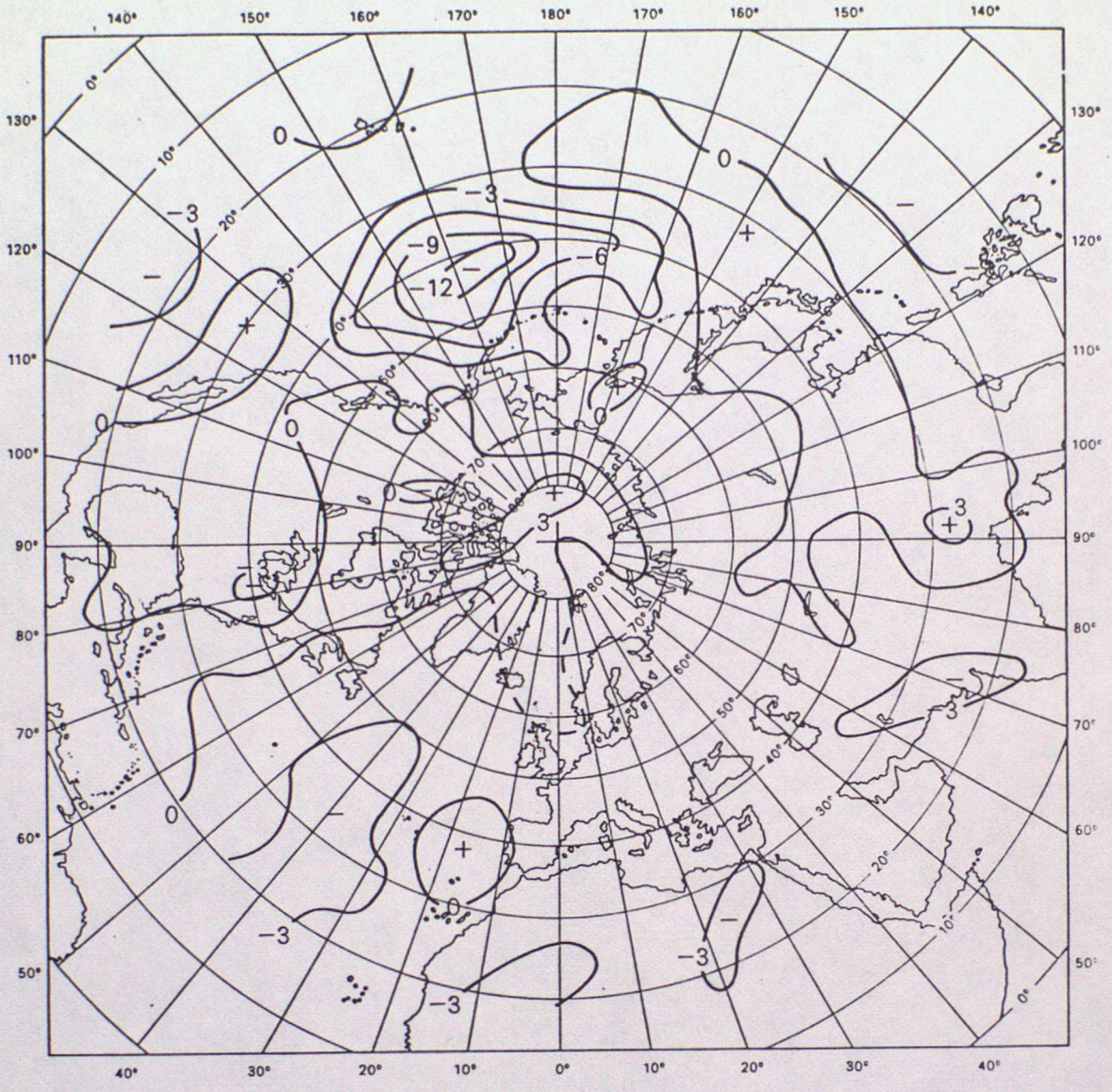


Figure 8. 500 mb geopotential difference, UK minus USSR, for January 1977, 00 GMT data. Contours at intervals of 3 decageopotential metres. (After Parker 1980)

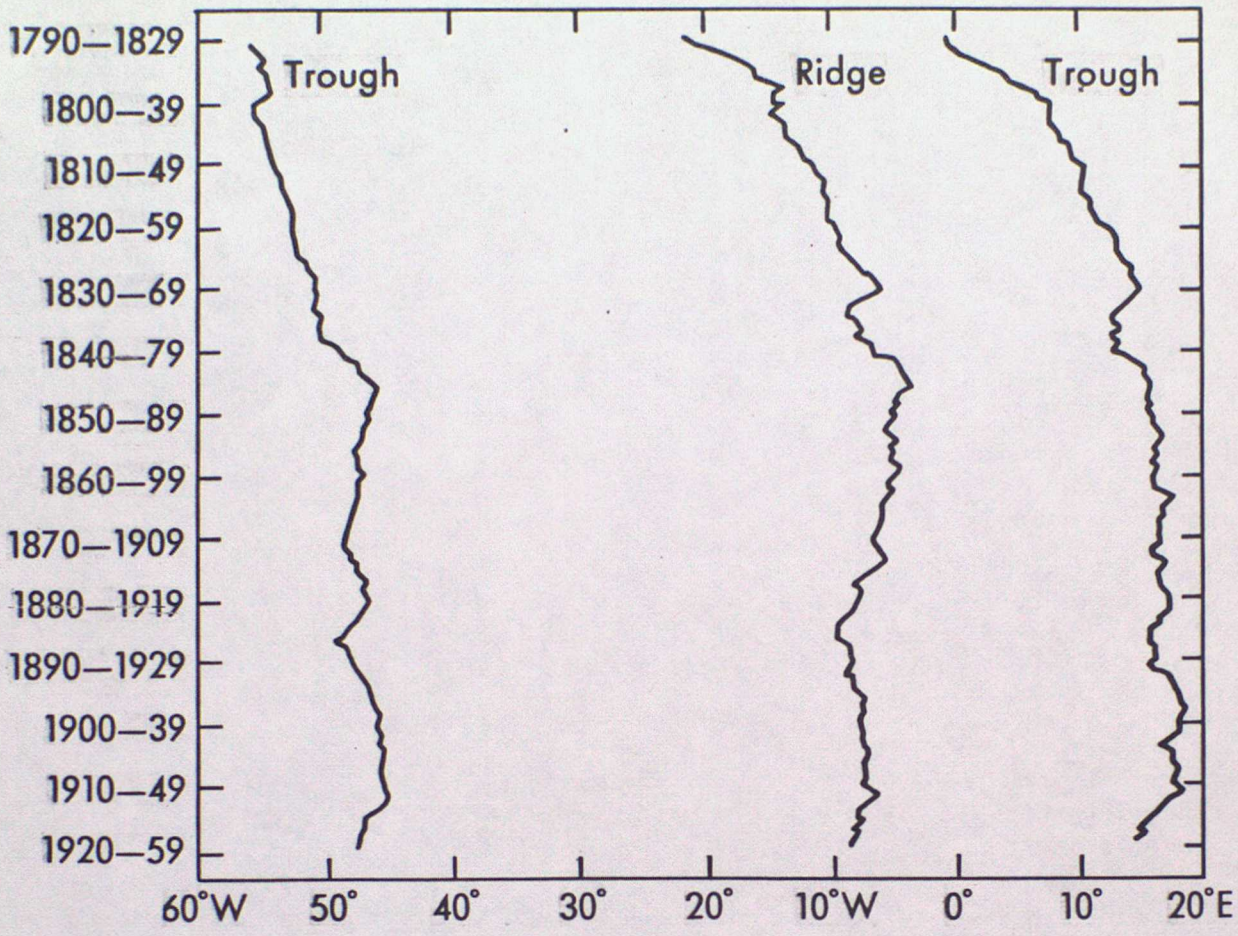


FIGURE 9. LONGITUDES OF SEMI-PERMANENT PRESSURE TROUGHS AND RIDGE (POSITIONS OF LOWEST AND HIGHEST AVERAGE PRESSURE AT MEAN SEA LEVEL) AT 45°N IN THE ATLANTIC SECTOR SINCE ABOUT 1800: JANUARY. (AFTER LAMB 1977). Forty-year running means.

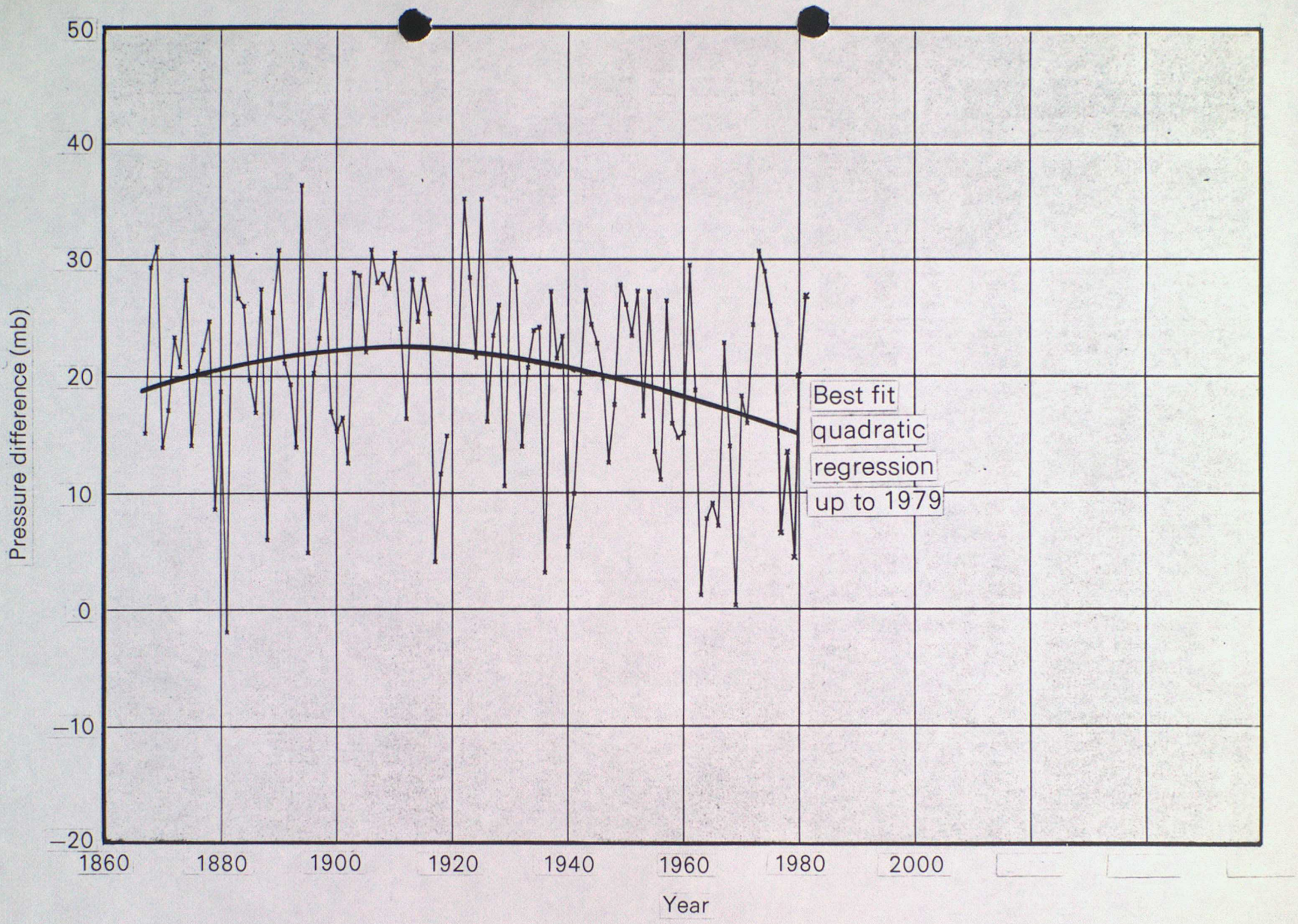
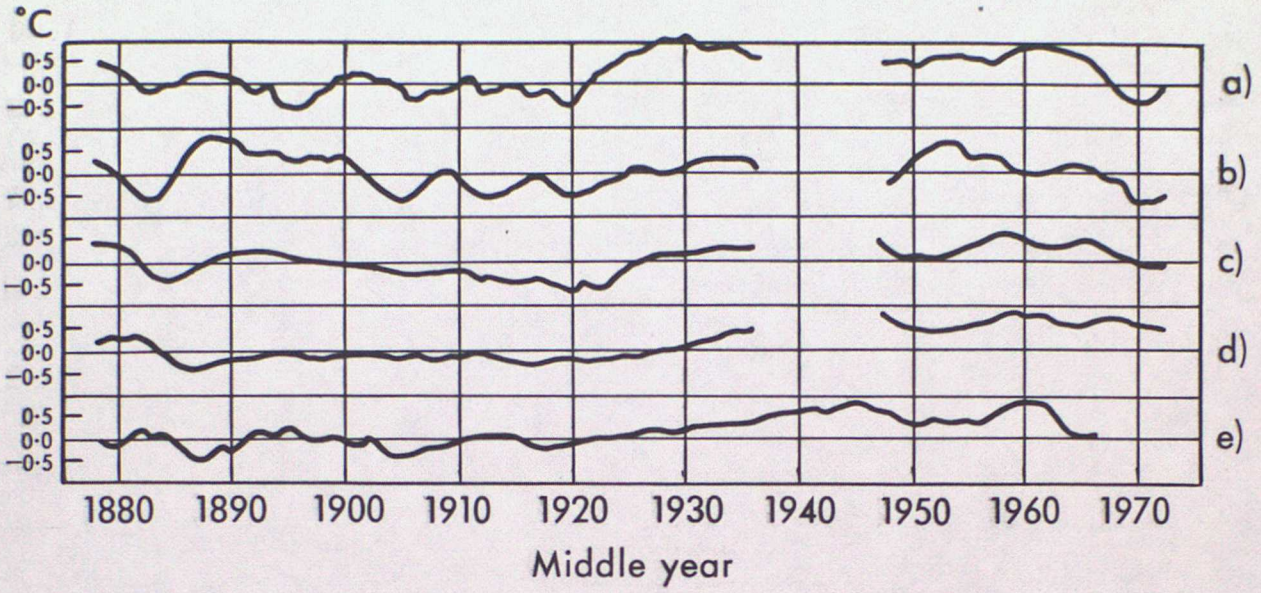


FIGURE 10. TIME SERIES OF AZORES (PONTA DELGADA) MINUS ICELAND (STYKKISHOLMUR) SEA LEVEL PRESSURE IN WINTER, 1867–1981.



LOCATIONS

a) 60—70°N, 50—60°W

b) 50—55°N, 30—50°W

c) 60—63°N, 20—30°W

d) 60—64°N, 0—10°W

e) TORSHAVN/MYGGENÆS HOMOGENISED SERIES (these island stations are within area d)

FIGURE 11. ANNUAL MEAN SEA SURFACE TEMPERATURE ANOMALIES IN THE NORTH ATLANTIC (AFTER SMED 1978) Values are 5-year running means.

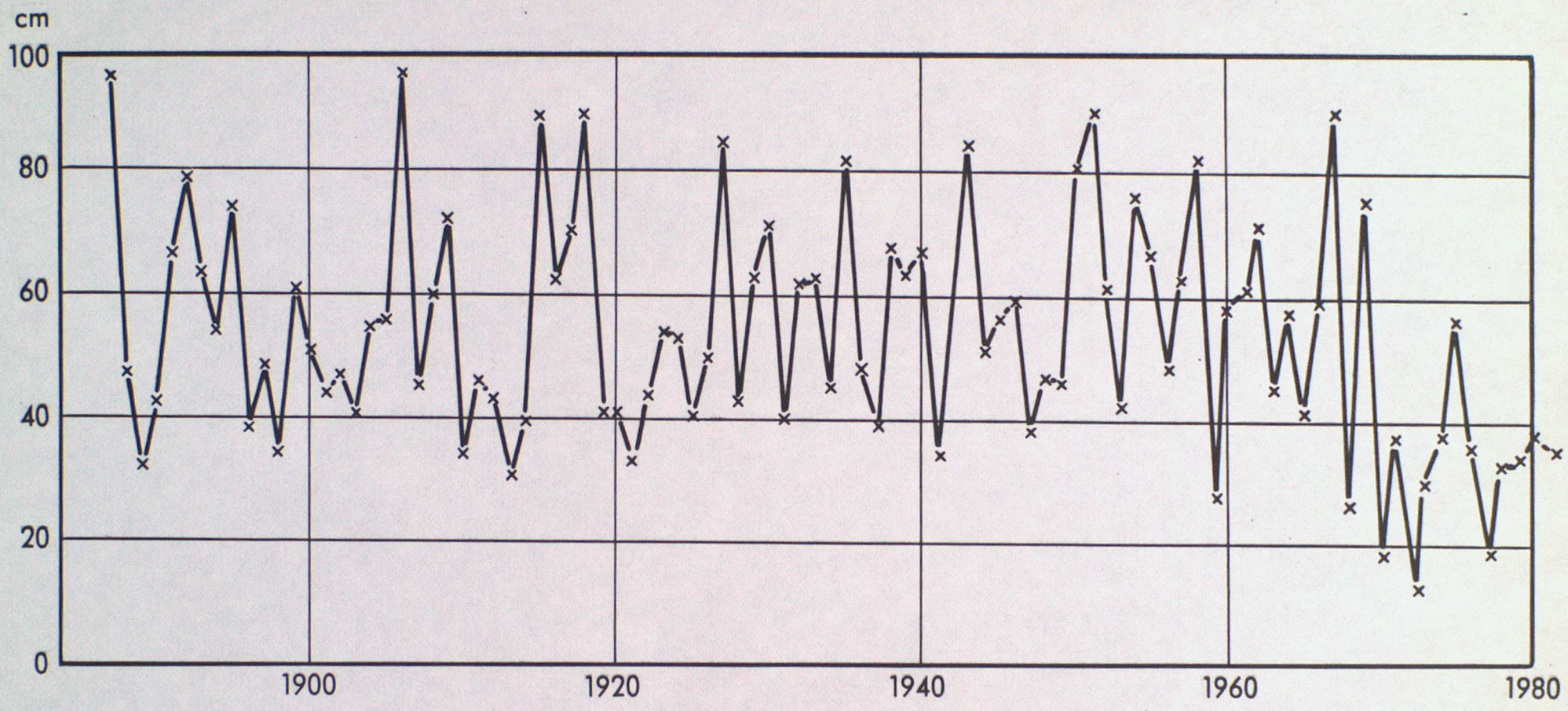


FIGURE 12. ANNUAL RAINFALL AT DAKAR, SENEGAL, 1887 TO 1981 (DATA UP TO 1972 ARE FROM LANDSBERG 1973)

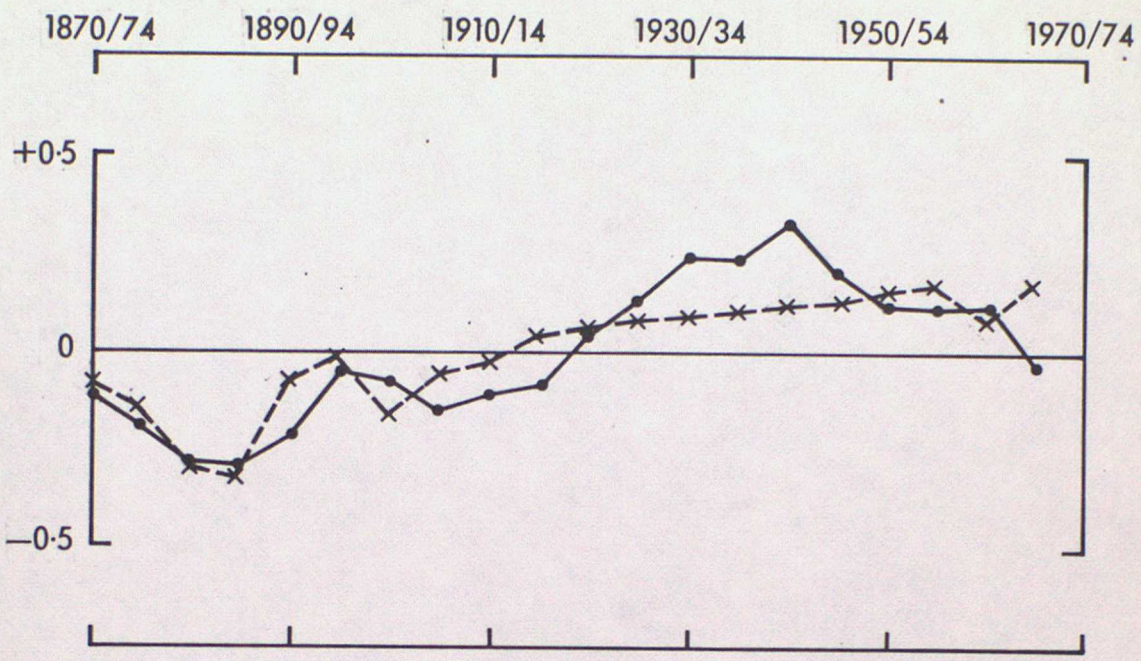


FIGURE 13. ACTUAL NORTHERN HEMISPHERE TEMPERATURE ANOMALIES (•——•) AND VALUES PREDICTED FROM MULTIPLE REGRESSION ON CO_2 AMOUNT, DUST VEIL INDEX, AND WOLF SUNSPOT NUMBER (x– –x) (AFTER MILES AND GILDERSLEEVES 1977). Values are 5-year means.